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INTRODUCTION

This second volume of contributions to the geology of the NWT comprises papers written by researchers under contract to the Geology Division of the Northern Affairs Program (NAP) in Yellowknife. Such papers were previously included with the NAP's annual Mineral Industry Reports, but the volume of material currently being generated justifies a separate publication. The articles in this volume were submitted to fulfill program-funded contracts, but we would like to take this opportunity to encourage submissions for Volume 3 from other sources, especially from industry geologists whose experience exploring for mineral deposits in the NWT would make their contributions particularly valuable. Submissions in either official language should be forwarded to W.A. Padgham, Chief Geologist, Northern Affairs Program, Box 1500, Yellowknife, NWT, X1A 2R3.

Ce deuxième volume des contributions à la géologie des TNO comprend des études écrites par les chercheurs sous contrat au Programme des Affaires du Nord (PAN) en poste à Yellowknife. Ces études étaient auparavant incluses dans les Rapports de l'Industrie du Minéral publiés annuellement par le PAN. Cependant, le volume de matériel présentement généré justifie une publication séparée. Nous voulons profiter de cette occasion pour encourager tous les chercheurs, spécialement les géologues de l'industrie privée, à soumettre des articles de façon à compléter un troisième volume. Les articles peuvent être soumis dans l'une ou l'autre des langues officielles à W.A. Padgham, Géologue en Chef, Programme des Affaires du Nord, C.P. 1500, Yellowknife, TNO, X1A-2R3.

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CHEMICAL COMPOSITION OF THE SPARROW PLUTON AND ASSOCIATED PEGMATITE DIKES

by

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ABSTRACT

The Sparrow Pluton, which is 8 km in diameter and lies 30 km northeast of Yellowknife, is one of a group of muscovite-bearing granite plutons emplaced in metasedimentary rock of the Yellowknife Supergroup. Determinations of major elements and some trace elements (Li, Zn, Rb, Sr, Zr, Nb, Ba) are presented for several samples of the granite and of various associated rocks. Pegmatite dikes occur throughout the pluton and within the schist to the east, and analyses of pegmatite from 3 dikes in granite and 3 in schist are also presented. The chemical data show that the sequence 1) central granite, 2) marginal granite, 3) granite dikes, 4) pegmatite, is one of decreasing Ti, Fe, Mg, Zr, and Ba. Pegmatite dikes in granite are rich in K and those in schist are rich in Na, while both are rich in P and Rb relative to the Sparrow granite. These data identify the central region of the Sparrow pluton or similar granite elsewhere in the associated plutons as a possible source for the pegmatite dikes.

INTRODUCTION

The Archean terrane near Yellowknife consists mainly of metamorphosed volcanic and sedimentary rocks, large masses of granitic rock dominated by granodiorite, and a group of smaller bodies of muscovite-biotite granite including the Duncan (1), Sparrow (2), and Prosperous (3) plutons (Fig. 1). The smaller plutons and the surrounding cordierite-zone metasedimentary rocks were invaded by countless pegmatite bodies, some of which contain beryl, spodumene and other rare-element minerals.

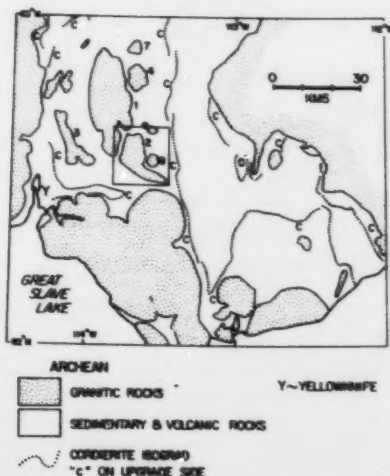


FIGURE 1: Simplified map of the Yellowknife region showing muscovite-bearing granite plutons northeast of Yellowknife and the location of Fig. 2 (from Henderson 1985).

The muscovite-bearing plutons were previously described by Henderson and Jolliffe (1941), Jolliffe (1942, 1946), Boyle (1961), Kretz (1968), Green and Baadsgaard (1971), Drury (1979), Kretz and others (1982) and McIntzer and others (1984), and the pegmatite bodies by Jolliffe (1944), Hutchinson (1955), Kretz (1968, 1970), and McIntzer and others

(1984 and this volume). Presented here are structural, mineral, and chemical data on the Sparrow pluton and on six pegmatite dikes near its eastern contact. These data may be helpful in defining the physical and chemical processes that gave rise to the granitic bodies and the associated pegmatite dikes.

METHODS

Pegmatite dikes were mapped by the chain-offset method. Mineral proportions within dikes were estimated visually with the aid of a Na cobaltinitrite stain to distinguish K-feldspar from albite. A strip of paper, 5 cm wide, was placed on the rock surface and was moistened with hydrofluoric acid. After 2-3 minutes, the paper was removed, the rock surface was rinsed with water, and a concentrated solution of Na cobaltinitrite was applied, which produces a yellow coating on K-feldspar but not on albite. The coating is removed by weathering in less than 10 years. Mineral proportions in granite were obtained by point-count analysis, using stained slabs of rock and thin-sections.

Standard procedures were followed to obtain chemical analyses. Samples of pegmatite and granite consisted of large fragments (500-1100 g) and channel samples. Powders were fused and analysed by x-ray fluorescence spectroscopy (XRF) at the University of Ottawa using a Phillips PW 1410/2 AHP instrument. Some powders were analysed for Na, K, and Li by atomic absorption (AA) analysis, also at the University of Ottawa. The $Fe^{2+}:Fe^{3+}$ ratio was not determined, and FeO in this report refers to total iron expressed as FeO.

THE SPARROW PLUTON

General Features

The Sparrow granite is a tadpole-shaped mass (Fig. 2), the body of which is here referred to as the Sparrow pluton and the tail as the Sparrow extension. This report deals only with the pluton. The granite consists of quartz, plagioclase (An 10), K-feldspar, muscovite and biotite in approximate ratios 34:35:18:9:3. Thus the plagioclase: K-feldspar ratio is approximately 2:1 and terms granite or granodiorite may be used (Streckeisen, 1973).

Apart from the narrow (1 km) connection with the

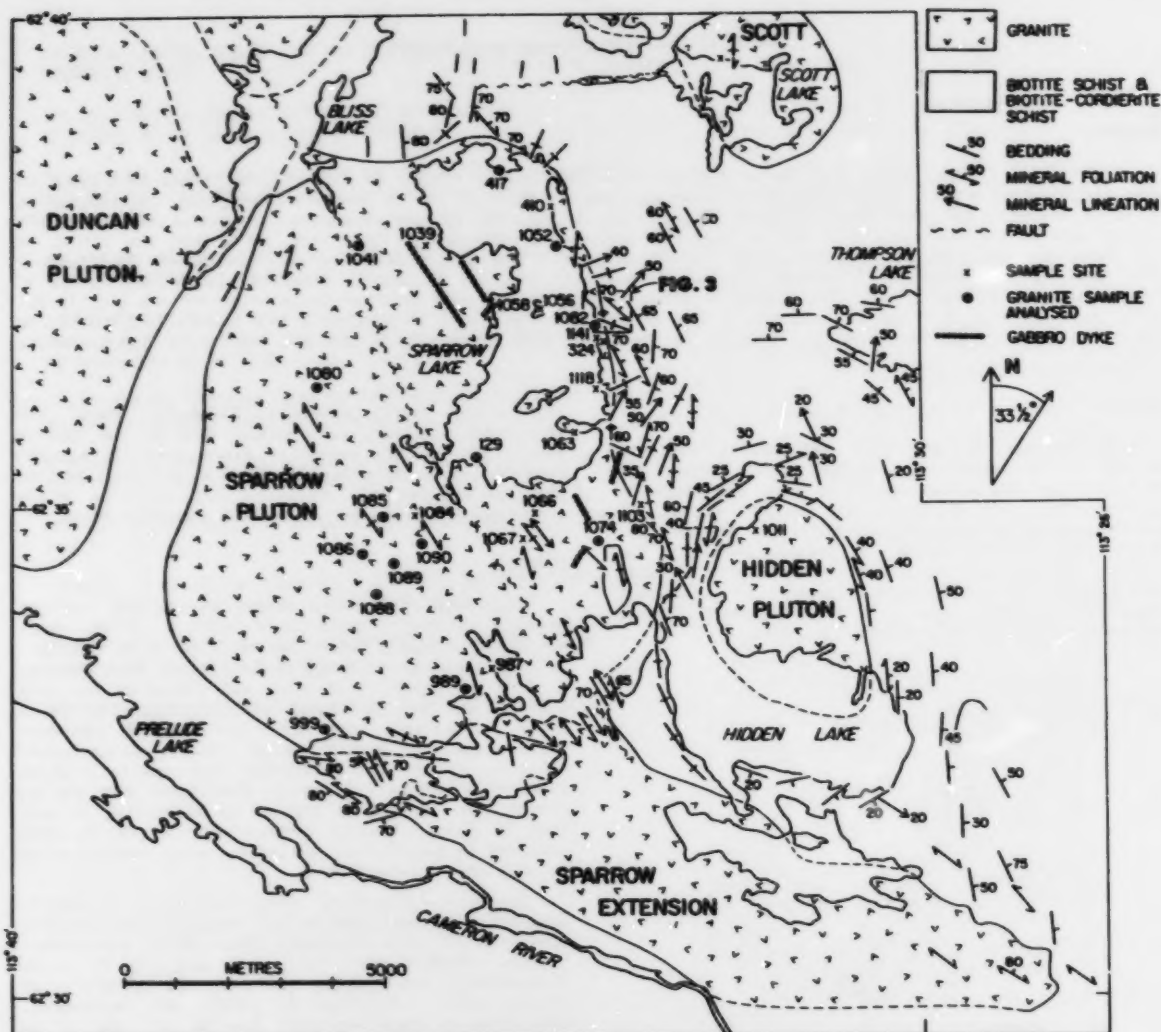


FIGURE 2: The Sparrow pluton; location of Fig. 3 is indicated.

Sparrow extension, the pluton is entirely surrounded by metasedimentary rock, chiefly biotite schist and biotite-cordierite schist. To the east of the pluton, these rocks were examined in detail by Kaminen (see Ramsay and Kaminen, 1977) who described several rocks consisting of different combinations of quartz, plagioclase, biotite, muscovite, cordierite, garnet, andalusite, sillimanite, cummingtonite, and gedrite. The important association quartz-muscovite-andalusite-sillimanite was observed in the eastern and southern contact zones, less than 100 m from the contact, and in inclusions in the margin of the pluton.

The structure of the country rocks near the pluton has not been examined in detail, but it is clear from Fig. 2 that to the east and north, all of

bedding, foliation, and lineation are truncated by the pluton. Foliation and lineation (both defined by crystals of biotite and cordierite) in relation to bedding and the vertical contact plane are also shown in Fig. 3. In contrast, all of these structures wrap about the smaller Hidden pluton (Fig. 2), which is richer in plagioclase and poorer in muscovite, and may be older.

Inclusions of schist, often angular, occur in the margin of the pluton; the largest of these is shown in Fig. 2. In the southern margin, some of the inclusions show isoclinal folds and axial plane foliation, indicating considerable deformation prior to granite intrusion. One fragment of schist, 30 m across, was observed east of site 1085, near the

Most of the granite is lightly porphyritic, with scattered crystals of K-feldspar up to 1 cm across. The microstructure is complex and variable; for example muscovite occurs as tabular crystals, 2-5 mm across, as fine-grained aggregates, and as oriented inclusions in plagioclase. Crystal orientation is apparently random, but some exposures, a few metres across, show a foliation and lineation defined by the parallel orientation of mica and feldspar crystals (e.g. at site 1067, Fig. 2). Mica foliation and lineation are more prominent in the southern margin of the pluton where the foliation lies nearly parallel to a regional north-to-northwest foliation (Fyson, 1975).

Granite dikes or apophyses project from the

pluton at several places (e.g. Fig. 3) and some of them (e.g. site 1103, southeast of Sparrow Lake) are conspicuously folded. The dikes are poor in K-feldspar and biotite and rich in muscovite, and are more heterogeneous than granite within the pluton. Pegmatite and quartz-tourmaline pods are abundant within some of them.

At places the granite-schist contact is sharp to the extent that it can be examined in a thin-section. Here the granite within a few cm or tens of cm from the contact decreases sharply in plagioclase and K-feldspar and increases in quartz and muscovite. Tourmaline may be present in this contact granite and in the adjacent schist.

Inclusions of a mica-rich rock are common throughout the pluton. These range in dimensions from a few cm to several metres and in composition from mica-rich granite to a rock composed predominantly of muscovite and biotite. The shape of these objects is extremely variable, as shown in Figs. 4b to 4f. They are well exposed along the southeast shore of Sparrow Lake in the vicinity of site 1118 (Fig. 2).

Patches of red alteration occur locally throughout the pluton, especially near the fault shown in Fig. 2. Numerous fracture sets were observed, the most prominent of which is a near-

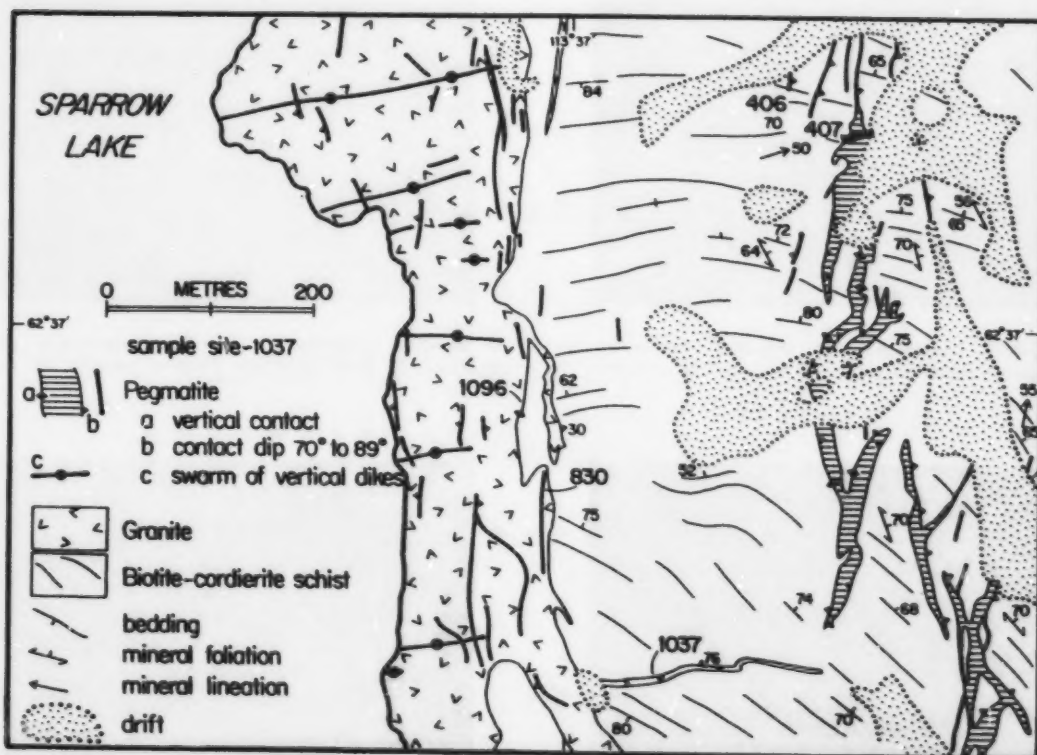


FIGURE 3: Eastern contact of the Sparrow pluton, showing pegmatite dikes in granite and in schist; locations of pegmatite sites 830 and 408 are indicated (from Kratz, 1968).

FIGURE 4: Inclusions in eastern margin of Sparrow pluton. Location of Fig. 4a is site 417 (Fig. 2), remainder in vicinity of site 1118 (Fig. 2).



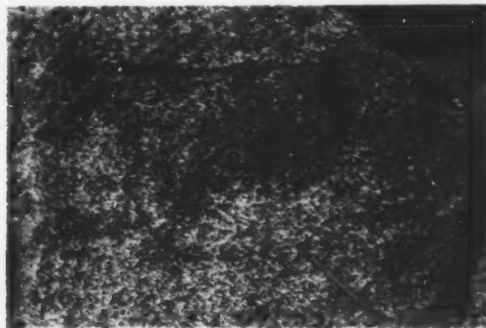
a) Inclusions of darker granite in lighter granite; contact (at pencil point) accentuated by ink line on rock surface.



b) Large mica-rich inclusion in granite, cut by pegmatite dike, and a small inclusion to left of pencil.



c) Angular inclusions in granite; inclusion at pencil point apparently partly replaced by granite.



d) Angular inclusion in granite, apparently partly replaced by granite, with original margin at pencil point.



e) Mica-rich inclusions in granite, cut by pegmatite dike.



f) Mica-rich inclusion in granite, cut by pegmatite dike (pencil on contact).

TABLE 1: SPARROW GRANITE

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
SiO ₂	73.7	73.7	73.9	74.3	73.5	73.5	73.2	74.6	73.8	73.6	74.2	75.5	73.6	73.8	74.6	74.3	73.6	73.9	75.1
TiO ₂	0.12	0.12	0.11	0.11	0.13	0.10	0.11	0.16	0.17	0.14	0.13	0.13	0.15	0.14	0.13	0.14	0.14	0.14	0.03
Al ₂ O ₃	15.2	14.9	14.4	14.5	14.5	14.5	15.6	14.4	14.7	14.9	14.2	13.2	14.5	14.6	14.2	14.8	14.6	13.7	14.8
FeO	1.09	0.98	1.06	0.99	1.08	1.02	0.61	1.18	1.30	1.05	1.33	1.30	1.10	1.27	1.12	0.86	1.13	1.41	0.63
MgO	0.10	0.10	0.11	0.11	0.12	0.05	0.00	0.22	0.19	0.23	0.20	0.22	0.21	0.16	0.07	0.07	0.14	0.21	0.00
MnO	0.02	0.02	0.02	0.02	0.03	0.03	0.02	0.02	0.02	0.02	0.02	0.01	0.02	0.02	0.02	0.02	0.03	0.02	0.01
CaO	0.63	0.60	0.60	0.62	0.64	0.60	0.62	0.66	0.66	0.60	0.61	0.46	0.70	0.63	0.58	0.60	0.62	0.65	0.47
Na ₂ O	3.66	3.73	4.05	3.76	3.63	4.14	3.27	3.37	3.45	3.48	3.20	2.82	3.47	3.79	3.46	3.33	3.59	3.20	4.83
K ₂ O	4.64	4.61	4.54	4.77	4.75	4.85	4.75	4.75	5.30	4.81	5.32	5.15	5.03	4.68	4.78	5.09	4.93	4.75	4.07
P ₂ O ₅	0.34	0.35	0.21	0.39	0.34	0.31	0.40	0.29	0.28	0.35	0.37	0.29	0.29	0.25	0.31	0.33	0.39	0.42	0.38
Li	215	220	200	200	215	99	118	68	106	45	160	77	52	150	110	110	220	140	17
Zn	60	64	59	51	54	49	28	40	65	32	51	43	50	64	55	48	77	58	35
Pb	290	290	280	270	280	270	270	190	260	230	220	160	220	290	260	270	320	180	180
Sr	50	49	73	69	68	71	35	99	83	62	94	89	84	75	63	67	67	77	25
Zr	56	54	68	69	68	66	42	83	95	66	82	76	86	93	67	66	71	71	21
Nb	9	11	7	8	9	4	8	11	7	9	9	5	9	4	9	10	10	10	7
Ba	220	210	310	310	320	340	140	430	410	290	450	440	400	320	270	310	300	350	86

1-5 Samples 1082-1 to 1082-5 obtained at intervals in shore-line distance of 50 m at site 1082 (Fig. 2).

6 Sample 417-3B, major part of granite at site 417 (Fig. 2).

7 Sample 417-3A, pod of darker granite at site 417 (Fig. 2).

8-12 Samples from central part of pluton at sites 1086, 1090, 1086, 1089, 1088 respectively (Fig. 2).

13-19 Samples at various places, from north to south, at sites 1041, 1052, 1080, 129, 1074, 989, 989 respectively (Fig. 2).

TABLE 2: ASSOCIATED ROCKS

	1	2	3	4	5	6
SiO ₂	74.6	74.3	74.5	73.6	71.1	58.9
TiO ₂	0.22	0.14	0.13	0.13	0.35	1.21
Al ₂ O ₃	15.1	14.6	14.7	14.5	15.1	18.1
FeO	1.75	1.03	1.04	1.19	2.90	9.05
MgO	0.33	0.10	0.06	0.14	0.67	2.40
MnO	0.03	0.02	0.02	0.02	0.05	0.09
CaO	0.78	0.55	0.57	0.54	0.67	0.43
Na ₂ O	1.31	3.46	3.55	4.05	2.26	0.55
K ₂ O	3.83	5.33	5.22	4.93	4.87	6.94
P ₂ O ₅	0.54	0.32	0.44	0.25	0.34	0.38
Li	210	150	180	138	440	1430
Zn	81	57	54	62	170	430
Rb	240	310	310	300	360	800
Sr	73	63	68	71	47	17
Zr	66	64	71	82	110	234
Nb	12	10	7	7	19	50
Ba	310	310	310	310	230	150

1-3 Granite at contact (1), 1.1 m from contact (2), and 2.3 m from contact (3), site 1096, (Fig. 3).

4 Fine-grained granite inclusion, site 1066 (Fig. 2).

5 Mica-rich inclusion, site 1144 (Fig. 2); includes Cr 10 ppm.

6 Mica-rich inclusion, site 1063 (Fig. 2); includes Cr 38 ppm.

horizontal set that has locally produced a stepped topography. The pluton is cut by a few vertical diabase dikes that are composed principally of augite, plagioclase, and magnetite, and range in width from 1 to 15 meters.

Chemical Composition

Estimates of the chemical composition of what appears to be typical Sparrow granite and an estimate of local variability are provided by 5 analyses listed in Table 1 (analyses 1 to 5), obtained at site 1082, and the average of these as listed in Table 3, column 1. This rock, with about 34 percent quartz, 35 percent plagioclase, 18 percent K-feldspar, 9 percent muscovite, and minor biotite (Table 4) is referred to as granite A. Small but detectable differences in composition exist within the 50-m-long area from which the five samples were obtained. Taking into consideration the uncertainty in each element determination as expressed by confidence limits (Table 5), it is evident that most pairs of analyses are indistinguishable with regard to most elements, but detectable variation does exist with regard to Al, Fe, Na, and possibly Sr and Ba.

At site 417 (Fig. 2), 3 km to the north, the dominant rock is a slightly lighter, mica-poor granite that contains bodies of a darker, mica-rich granite as shown in Fig. 4a. Analyses of the lighter and darker granite appear as analyses 6 and 7 in Table 1. Chemically, the lighter rock is very similar to granite at site 1082, described above, and is viewed as a variant of granite A. The slightly higher Na content is related to a slightly higher proportion of plagioclase (Table 4). The conspicuously higher K-feldspar and lower muscovite have not affected a change in K content, but the rock presumably contains less H₂O.

The darker granite at site 417 (analysis 7, Table

TABLE 3: VARIETIES OF GRANITE AND PEGMATITE

	1	2	3	4	5	6	7	8
SiO ₂	73.8	73.7	73.3	73.2	75.1	75.3	76.7	71.8
TiO ₂	0.12	0.10	0.15	0.11	0.03	0.01	--	--
Al ₂ O ₃	14.7	14.7	14.3	15.6	14.8	14.8	13.0	16.1
FeO	1.04	1.00	1.23	0.61	0.63	0.55	0.10	0.21
MgO	0.11	0.07	0.21	0.00	0.00	0.00	--	--
MnO	0.02	0.02	0.02	0.02	0.01	0.02	--	--
CaO	0.62	0.61	0.60	0.62	0.47	0.28	0.37	0.66
Na ₂ O	3.77	4.12	3.26	3.27	4.83	4.68	2.3	6.0
K ₂ O	4.66	4.83	5.07	4.75	4.07	3.59	7.4	3.1
P ₂ O ₅	0.35	0.32	0.32	0.40	0.38	0.54	0.8	1.0
Li	210	96	90	118	17	17	14	30
Zn	58	49	46	28	35	70	20	30
Rb	280	270	210	270	180	420	800	470
Sr	62	73	85	35	25	4	30	90
Zr	63	68	80	42	21	18	--	10
Nb	9	4	8	8	7	16	2	9
Ba	274	330	400	140	86	64	80	70

1. Granite A, mean of 5 samples at site 1082 (analysis 1-5, Table 1).

2. Granite A, mean of 2 samples at site 417 (including analysis 7, Table 1).

3. Granite A, mean of 5 samples at center of pluton (analyses 8-12, Table 1).

4. Granite B, mean in granite A at site 417 (Fig. 4a).

5. Granite B, south margin of pluton, site 999 (analysis 19, Table 1).

6. Granite B, dike at site 1037 (Fig. 3).

7. Pegmatite, estimate of mean composition of dike A, site 830 (Fig. 8).

8. Pegmatite, estimate of mean composition of dike A, site 406 (Fig. 9).

1) which in terms of mineral proportions is very similar to granite at site 1082 (Table 4) is higher in Al and conspicuously low in Fe, Mg, Sr and Ba (Column 4, Table 3). This will be referred to as granite B.

Five analyses from the center of the pluton, in an area 2 X 2 km, appear as analyses 8 to 12 in Table 1. These are identified as granite A, but compared with the rock at site 1082, mean Fe, Mg and Ba are high and Rb is low (Column 3, Table 3).

Several additional analyses, obtained at various places across the pluton are listed as analyses 13 to 19 in Table 1. Samples 13 to 18 resemble granite A as defined above, but the range in composition for most elements is greater, as would be expected. Analysis 19 (site 999 in the south margin of the pluton) shows low Fe, Mg and Ba and forms an example of granite B (Table 3). With regard to mineral proportions (Table 4), this rock is not unique, and indeed it is evident that the distinction between granite A and B must be based on chemical composition alone.

An analysis of a specimen of granite from the dike shown in Fig. 3 (site 1037) appears as column 6 in Table 3. This rock is low in Fe, Mg, Ba and is therefore similar to granite B. It is low in Li and high in Rb.

The elements Fe and Ba may be chosen to discriminate granites A and B, as shown in Fig. 5. This figure also shows a tendency for granite centrally located in the pluton to be richer in Fe and Ba than granite in the margin.

An analysis of the quartz-muscovite rock found at the granite contact (site 1096, Fig. 3) and of granite 1.1 and 2.3 meters from the contact are listed in Table 2. Assuming that the contact rock was initially composed of granite as presently found 1-2.5 m from the contact, the change in composition that occurred resulted in an increase in Al, Mg, Fe

TABLE 4: APPROXIMATE MINERAL PROPORTIONS*

	Qtz	Pl	Kf	Ms	Bt	Ap	Trm
Granite A (1082)	34	35	18	9	3	<1	-
Granite A (417)	30	38	26	2	4	<1	<1
Granite B (417)	30	33	23	10	3	<1	-
Granite B (999)	34	37	22	5	1	<1	-
Sparrow pluton (mean of 19)	32	31	24	12			
Granite dike (1037)	34	45	9	12	-	<1	<1
Granite at contact (1096)	54	5	<1	30	1	<1	<1
Granite 1.1 m from contact	32	35	20	10	2	<1	-
Mica-rich inclusion (1141)	25	25	16	22	11	<1	-
Mica-rich inclusion (1063)	32	1	1	29	35	1	<1

* Qtz quartz, Pl plagioclase, Kf K-feldspar, Ms muscovite, Bt biotite (partly altered to chlorite), Ap apatite, Trm tourmaline.

TABLE 5: ANALYTICAL UNCERTAINTY

	V	±C		V	±C
SiO ₂	1	1.3	Li	3	12
TiO ₂	1	0.004	Zn	10	10
Al ₂ O ₃	1	0.24	Rb	40	80
FeO	1	0.016	Sr	4	3
MgO	10	0.05	Zr	10	7
MnO	10	0.004	Nb	25	5
CaO	2	0.03	Ba	10	30
Na ₂ O	2	0.14			
K ₂ O	1	0.12			
P ₂ O ₅	4	0.03			

V = coefficient of variation

±C = confidence limits to numbers in Table 1

TABLE 6: ESTIMATED MEAN MINERAL PROPORTIONS IN PEGMATITIC DIKES

	Qtz	Ab	Kf	Ms
Dikes in granite (830)				
Dike A (9)	15	45	30	10
Dike B (3)	15	45	40	5
Dike C (2)	25	30	40	5
Dikes in schist (406)				
Dike A, margin (10)*	20	70	≤1	10
Dike A, center (2)	25	30	40	10
Dike B, margin (22)	30	60	≤1	10
Dike B, center (5)	20	10	60	10
Dike B, filament (10)	35	65	≤1	≤1
Dike C (2)	20	70	≤1	10

* In brackets, number of places where estimates were obtained.

and Ca, and a decrease in K and Na. All other elements remained virtually unchanged, which suggests that little change in volume occurred. Therefore the enriched elements were added and the depleted were removed, possibly by exchange of ions with the adjacent schist (Fig. 6).

One analysis of a fine-grained granite inclusion is listed in Table 2, analysis 4. This rock is virtually identical to granite A at site 1082.

Two kinds of mica-rich inclusions appear to be present in the Sparrow pluton. Some inclusions form angular fragments (e.g. at pencil point in Fig. 4c and 4d) which were evidently partly to nearly completely replaced by granite. Others are spherical shells of mica-rich material (Fig. 4e and 4f), presumably formed by the segregation of muscovite and biotite, i.e. by the removal of quartz and feldspar from granite. Scattered fragments of mica-rich material may represent broken pieces of shells, such as those shown in Fig. 4e.

Chemical data were obtained on two mica-rich

TABLE 7: DIKES IN GRANITE (SITE 830)

Dike:	A			B		C
	1	2	3	4	5	6
SiO ₂	66.9	68.2	76.9	76.7	66.0	87.3
Al ₂ O ₃	18.9	17.7	13.0	13.0	18.1	6.6
FeO	0.27	0.17	0.25	0.10	0.80	0.50
CaO	1.0	0.43	1.0	0.37	0.72	0.14
Na ₂ O	7.6	3.6	5.0	2.3	7.7	1.3
K ₂ O	2.9	8.2	2.2	7.4	3.2	2.3
P ₂ O ₅	1.3	0.9	1.0	0.8	1.3	0.2
Zn	12	70	30	20	31	72
Rb	270	1200	180	800	250	180
Sr	78	32	35	30	46	24
Zr	11	--	7	--	8	8
Nb	9	26	9	2	4	3
Ba	94	110	85	80	70	130

1. Margin, in contact with schist, sample 19.
2. Margin, in contact with granite, sample 22.
3. Centre, sample 20.
4. Wall-to-wall, sample 40; includes 14 ppm Li.
5. West half of dike, sample 23.
6. Wall-to-wall, sample 25.

fragments of the second kind. Mineral proportions are listed in Table 4 and the analyses in Table 2 (analyses 5 and 6).

The relation of the mica-rich inclusions to granite A is shown in Fig. 7. Linear trends appear to exist for several elements when plotted against SiO₂, and most of these, when projected to higher concentrations, meet the horizontal axis at 76 percent SiO₂. These relations indicate that it would be possible to obtain mica-rich inclusions, progressively more mafic, by extracting from granite A variable amounts of a substance composed of 76 percent SiO₂, 14 percent Al₂O₃, and somewhat variable amounts of Na₂O and K₂O.

PEGMATITE BODIES

General Features

Pegmatite dikes composed principally of quartz, albite, K-feldspar, and muscovite occur throughout the Sparrow pluton. Dike width ranges from 1 cm to

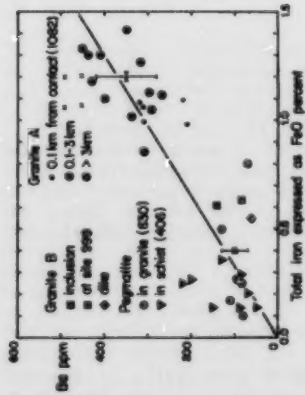


FIGURE 5: Fe and Ba content of granite and pegmatite; length of error bars indicated approximately.

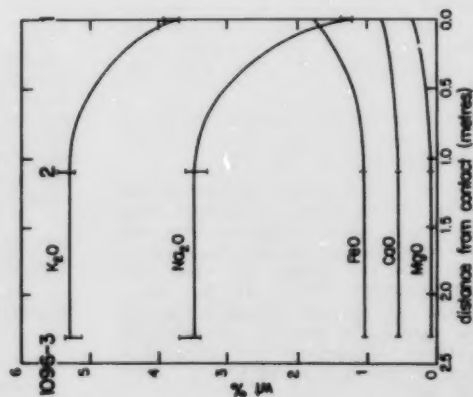


FIGURE 6: Loss of alkalis and gain of Fe, Ca, and Mg at granite contact, site 1080, Fig. 3.

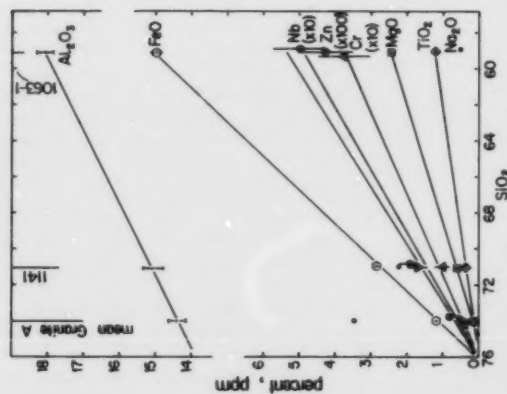


FIGURE 7: Variation diagram to compare mica-rich inclusions (1141 and 1063-1) with Sparrow granite (A).

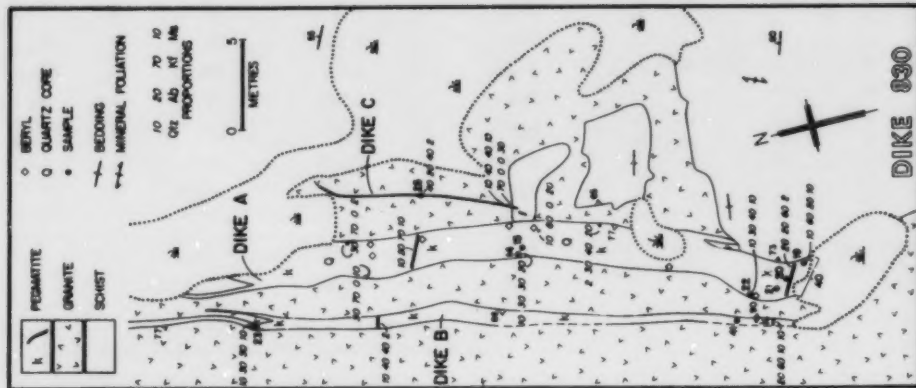


FIGURE 8: Pegmatite dikes in granite at site 830 (Fig. 3); mineral symbols are Qtz quartz, Ab albite, Kf K-feldspar, Ne muscovite.

several metres, and the orientation varies greatly. However, locally in the east margin of the pluton, swarms of narrow dikes strike easterly and dip vertically, as shown in Fig. 3.

Dikes similar in composition but more irregular in shape occur within the metasedimentary rocks to the east (Fig. 3), extending as far as the cordierite isograd shown in Fig. 1 (Fortier, 1947). Compared with very narrow dikes in granite, which invariably contain K-feldspar, those in schist are commonly free of K-feldspar.

A zonal pattern exists with regard to tourmaline, beryl, and spodumene. Tourmaline is common in dikes within the pluton, beryl occurs in some dikes in the east margin of the pluton (e.g. site 1063, Fig. 2; site 830, Fig. 3), and in some dikes in schist extending as far as 8 km from the contact. Spodumene is confined to a zone extending from approximately 4 to 8 km from the pluton. Further information on these rocks may be found in a report previously published (Kretz, 1968).

Chemical Composition

Three pegmatite dikes in granite at site 830 (Fig. 3) and three in schist at site 406 (Fig. 3) were selected for detailed study. The purpose of the study is to compare pegmatite dikes in granite with those in schist, and to compare both with the Sparrow granite. Information on the geometry and mineral content of the dikes is presented in Figs. 8 and 9.

At site 830 (Fig. 8), dike A is a beryl-bearing pegmatite which varies in mineral content from place to place and locally contains a core of quartz. Dike B is similar to A (Table 4) but is free of beryl. Dike C, which is 3 cm wide, is rich in quartz. At the southern termination of this small dike, where it enters schist, it consists entirely of quartz and muscovite, i.e. a rock similar to the contact granite described above.

Chemical analyses of pegmatite from different parts of the three dikes are listed in Table 7. Concentrations of Ti, Mg, and Mn lie near or below the limit of detection and are not listed. An estimate of the mean composition of dike A is given in Table 3, column 7, where it may be compared with granite A and granite B.

Pegmatite dikes at site 406 in schist (Fig. 3) are shown in Fig. 9. The dikes are more irregular in shape than those described above, presumably the result of deformation. Dikes A and B are relatively small (Fig. 3) while dike C is very small, only 9 m long.

Dikes A and B are composed of quartz, albite, K-feldspar, and muscovite, and in addition, garnet and tourmaline are locally present in B. Small dike C is evidently free of K-feldspar.

Dikes A and B are zoned; the marginal zone is virtually free of K-feldspar and the central zone contains both feldspar minerals. Very narrow dikes, less than 3 cm wide (here referred to as filaments) which branch out from B, are also lacking in K-feldspar and some of these contain only quartz and albite. Hence the mineral content of dikes in schist appears to be related to dike width (Table 6).

Chemical analyses of pegmatite at site 406 were obtained at numerous localities and the results are listed in Table 8. In comparing marginal and central zones (1 and 2; 4 and 5) we find that despite

differences in mineral content, the zones are very similar in chemical composition. This may be attributed to slightly higher muscovite in the margins, compared with the centers, where the additional K resides in K-feldspar. However, in filaments where neither muscovite nor K-feldspar are present, K is decidedly low (Table 8, Column 7) and the rock is exceptionally low in Rb and high in Sr.

In examining the data of Tables 7 and 8 it is clear that pegmatite at each of the two localities varies in composition, especially with regard to alkali and trace elements. The major compositional differences are that dikes in schist appear on average to be richer in Na while those in granite are richer in K.

When comparing pegmatite with granite (Table 3) it is clear that pegmatite is poorer in Ti, Fe, Mg, Mn, Li, Zr and Ba, and richer in P and Rb relative to granite A. Also, the Na and K content of pegmatite in volumes of 1000 cm³ or larger shows much more variation than in granite. This is shown in the Na-K plot of Fig. 10, based on XRF analyses, referred to above, supplemented by AA analyses of additional samples (Table 9). Plotted also in Fig. 10 are mean concentrations of Na and K in albite, K-feldspar, and muscovite of adjacent dike 407 (Fig. 3) as previously determined (Kretz, 1970). Note that the Na and K content deviates in various directions from that of the normal Sparrow granite (granite A).

DISCUSSION

An estimate of the mean composition of the Sparrow pluton is given in Table 10, column 1. This is based on analyses of granite A at 13 localities within the pluton (Fig. 2). Analyses of granite B, which presumably forms only a small proportion of the pluton by volume, have been omitted. A composite analysis of the smaller Prestige pluton to the north (pluton 7, Fig. 1) is listed in column 2 of Table 10. This is a single analysis of equal fractions of powders of 17 uniformly spaced samples from the pluton, thoroughly mixed. Information on this pluton was previously presented (Kretz and others, 1982). The two granite bodies are evidently very similar in composition with regard to all 17 elements that were determined, but the Prestige granite appears on average to be slightly richer in Fe, Mg, and K. A composite analysis of the Prosperous pluton to the west (pluton 3, Fig. 1) is listed in column 3 of Table 10. This granite is also similar to the Sparrow granite but is slightly richer in Na and poorer in K.

The results obtained for the Sparrow pluton do not reveal a conspicuous chemical zonation, but Figure 5 suggests that the broad central part of the pluton is richer in Fe and Ba than the narrow margins.

The present results for Li are interesting in relation to the occurrence of spodumene-bearing pegmatite bodies lying 4 to 8 km to the east. Li concentrations of granite A range from 45 to 220 ppm with a mean of 118 ppm. A granite dike projecting into schist contains 17 ppm Li. A chip sample across pegmatite dike in granite contains 14 ppm Li, and a pegmatite dike in schist contains an estimated 18 ppm Li (Tables 7 and 8). Hence the Li content of some pegmatite dikes in and near the pluton is much lower than in the pluton. By comparison the Li content of spodumene-bearing pegmatite dikes to the east is much higher; one of these was found to contain 1500 ppm Li

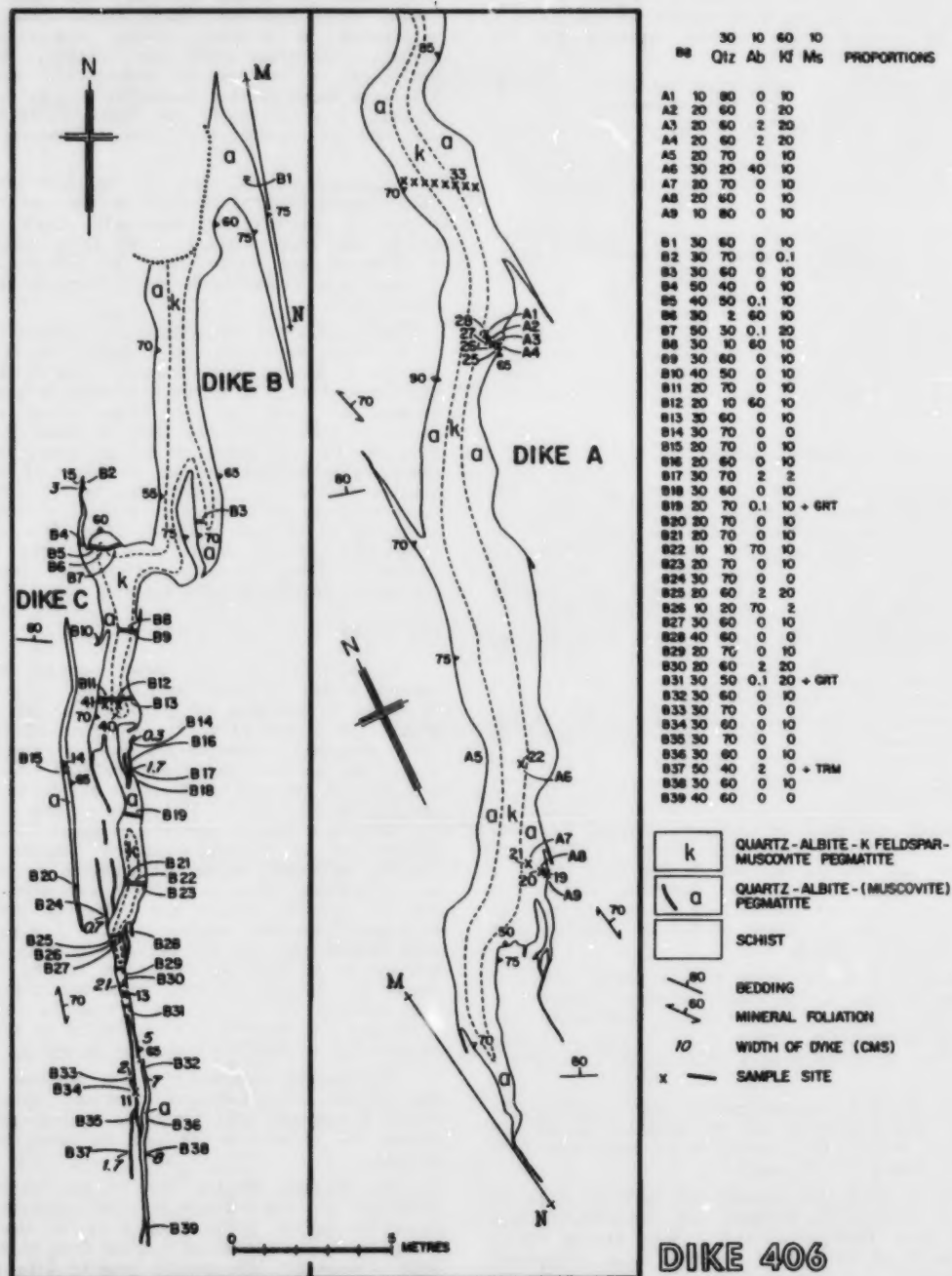


FIGURE 9: Pegmatite dikes in schist at site 406 (Fig. 3); relative position of dikes A and B is obtained by superimposing lines M-N; mineral symbols as in Fig. 8, plus Grt garnet, Trm tourmaline.

TABLE 8: DIKES IN SCHIST (SITE 406).

Dike:	A			B			C	
	1	2	3	4	5	6	7	8
SiO ₂	73.8	72.3	71.8	70.5	72.5	70.1	73.4	72.4
Al ₂ O ₃	15.6	15.3	16.1	17.3	15.5	18.3	15.3	17.2
FeO	0.29	0.14	0.21	0.14	0.14	0.38	0.27	0.25
CaO	0.62	0.85	0.66	1.0	1.4	0.36	1.41	0.25
Na ₂ O	5.10	5.8	6.0	8.2	7.4	6.7	6.7	5.1
K ₂ O	2.3	3.1	3.1	1.4	1.4	2.3	0.6	5.6
P ₂ O ₅	0.76	1.2	1.0	1.1	1.5	0.5	0.2	0.3
Zn	90	30	30	20	20	40	100	30
Pb	400	450	470	210	120	360	10	400
Sr	40	50	90	30	30	50	270	80
Zr	18	9	10	30	90	30	20	10
Hf	16	8	9	14	5	20	50	20
Ba	85	50	70	90	150	130	200	220

1. Marginal zone, mean of four samples (25,26,27,28).
2. Central zone, sample 29 (northern continuation).
3. Wall-to-wall, sample 33.
4. Marginal zone, sample 41; includes Li 23 ppm.
5. Central zone, sample 40; includes Li 12 ppm.
6. Wall-to-wall, sample 13, near termination of dike.
7. Filament, sample 15.
8. Wall-to-wall, sample 14.

TABLE 9: ALKALI CONTENT OF PEGMATITE, DETERMINED BY ATOMIC ABSORPTION (AA) AND X-RAY FLUORESCENCE (XRF) ANALYSIS.

	Na ₂ O		K ₂ O	
	AA	XRF	AA	XRF
406-11	5.7		0.83	
-13	6.4	6.7	2.1	2.3
-14	5.0	5.1	2.3	2.6
-15	6.5	6.7	0.70	0.58
-19	5.7		2.1	
-20	4.5		2.4	
-21	6.8		0.91	
-22	2.9		11.2	
-25	3.9	3.7	3.3	3.5
-26	4.6	4.6	2.6	2.8
-27	5.1	5.0	1.2	1.3
-28	7.0	7.0	1.4	1.4
-29	5.8		3.1	
-33	5.8	6.0	2.9	3.1
-40		7.4		1.4
-41		8.2		1.4
830-14	4.4		4.6	
-15	2.8		9.7	
-19	7.5	7.6	2.7	2.9
-20	5.0	5.0	1.9	2.2
-21	2.3		11.7	
-22	3.6	3.6	8.4	8.2
-23	8.0	7.7	3.0	3.2
-25	1.3	1.3	2.0	2.3
-40		2.3		7.4
-41		3.7		1.8

(Kretz, 1968).

The origin of the Sparrow pluton and associated pegmatite dikes will now be discussed very briefly, based on the newly acquired data. Assuming that emplacement of the pluton was a slow process, temperatures within the wall rocks and within schist inclusions must have been comparable to temperatures within the margin of the pluton. The association andalusite-sillimanite-muscovite-quartz, as found in schist inclusions in the southern margin of the pluton is particularly informative for it indicates a temperature near 600°C and a pressure near 2.5 kb, corresponding to a depth of about 9 km (Kretz and others, 1982). Under these conditions, solid muscovite granite is stable relative to completely or partially molten granite, dry or water-saturated (Huang and Wyllie, 1973).

However, the remarkable homogeneity of the Sparrow granite suggests that the granite passed through a molten stage. Thus, whether the granite was derived from Yellowknife Group metasediments as proposed by Green and Baadsgaard (1971), from the basement to these sediments or from deeper levels, it is suggested that a layer of silicate melt formed within the crust beneath the group of plutons as presently exposed. The layer then began to rise due to buoyant forces and to produce a number of domes, as proposed by Henderson (1943), Kretz (1968), Drury (1977) and Kretz and others (1982). As lower temperatures were encountered, the rock presumably solidified but continued to rise by plastic (ductile) flow, to temperatures of about 600°C. The low pressure recorded by the metasediments, corresponding to a depth of only 9 km, when compared to the large surface area of the Sparrow pluton, suggests that the domes spread out to form mushroom-shaped bodies, as produced in the experiments of Ramberg (1970). Many of the complex (non-igneous) textural features and possibly some of the variation in alkali content may be attributed to this latter stage of emplacement (Kretz and others, 1982).

The distribution of pegmatite within and about the muscovite-bearing granite plutons near Yellowknife suggests that either the pegmatite was derived from the granite or both have a common source. The chemical data here presented show that relative to normal Sparrow granite (A), the associated pegmatite dikes are poor in Ti, Fe, Mg, Mn, Li, Zr and Ba, and rich in P and Rb. Granite dikes and some masses of granite within the pluton (granite B) occupy an intermediate position in as much as they contain intermediate amount of Fe, are low in Mg and Ba, and show a tendency to be low in Ti, Li, Sr, Zr, and Ba, and high in P and Rb (Table 3).

Granite centrally situated in the Sparrow pluton, when compared to granite in the margin, appears to be slightly richer in elements that are low or absent in the pegmatite dikes, notably Ti, Fe, Mg, Zr and Ba. Also, Rb, which is enriched in pegmatite, is depleted in the central granite (Table 3). It is suggested therefore that the pegmatite was derived from the central granite or from similar granite which may occur within the associated plutons or at some depth beneath the group of plutons. According to the above model of granite emplacement, central portions of the Sparrow and other plutons represent deeper and presumably hotter material.

The observation that pegmatite dikes in the study area differ in mineral content and bulk composition indicates that pegmatite-forming matter of different composition was produced at the source region or

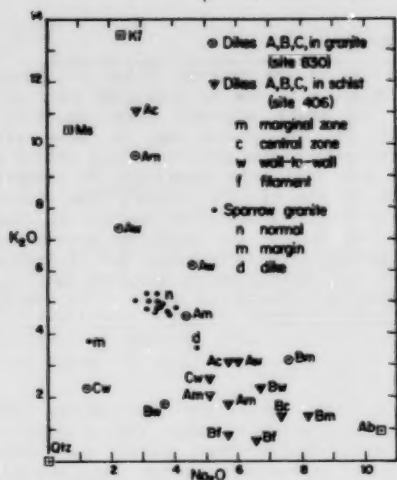


FIGURE 10: K and Na content of pegmatite compared with normal Sparrow granite (granite A) showing also K and Na content of K-feldspar (Kf), muscovite (Ma), and albite (Ab) in adjacent dike 407 (Fig. 3).

TABLE 10: MEAN SPARROW GRANITE COMPARED WITH PRESTIGE GRANITE AND PROSPEROUS GRANITE

	1	2	3
SiO ₂	74.1	72.8	74.3
TiO ₂	0.14	0.20	0.07
Al ₂ O ₃	14.4	14.8	14.4
Fe ₂ O ₃			0.32
FeO	1.16*	1.23*	0.77
MgO	0.16	0.27	0.16
MnO	0.02	0.01	0.04
CaO	0.62	0.62	0.60
Na ₂ O	3.47	3.39	4.18
K ₂ O	4.93	5.36	4.15
P ₂ O ₅	0.33	0.32	0.11
H ₂ O			0.42
Li	118	70	
Zn	53	54	
Rb	242	250	300
Sr	77	86	65
Zr	76	92	
Nb	8	9	
Ba	350	415	520

1. Sparrow granite A (mean of 13).
2. Prestige granite (composite).
3. Prosperous granite (composite, Boyle, 1961); trace elements from Drury (1979).

* Total iron expressed as FeO.

that the dikes were 'open systems' receiving by crystallization only some of the elements present in the pegmatite-forming material, the remainder moving on to crystallize elsewhere, presumably at lower temperatures. The latter interpretation is preferred for it also provides an explanation for the regional zonation of tourmaline, beryl and spodumene.

Concerning the nature of the pegmatite-forming matter, the dominant opinion at present is that pegmatite crystallizes from a silicate melt which contains H₂O, B, and other volatiles in solution (Jahns, 1982). Eastward from Sparrow Lake the pegmatite-forming matter has evidently migrated through metasediments along channels too narrow to be detectable and for distances of several km. Compared with mafic to felsic silicate melts including those saturated with water, the pegmatite-forming matter was evidently much less 'cohesive' and more mobile. The down-temperature migration of ions and ionic clusters, aided by the presence of H₂O and other volatiles, forms a valid alternative hypothesis.

More information is obviously needed on other granite and pegmatite bodies of the Yellowknife region before a satisfactory understanding of these interesting rocks may be obtained.

ACKNOWLEDGEMENTS

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2

FINAL REPORT ON THE GEOLOGICAL FIELD STUDY,
YK 83-84/008
THE YELLOWKNIFE PEGMATITE FIELD: MINERALOGY AND
GEOCHEMISTRY OF Nb-Ta-Sn OXIDE MINERALS

by

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INTRODUCTION

The pegmatites of the Yellowknife pegmatite field are known for their content of rare-elements such as Be, Nb, Ta, Sn and Li. Meintzer and others (1984) presented a summary of the mineralogical and structural aspects of pegmatites and fertile granites in the area. The present report provides a discussion of the distribution and geochemistry of Nb-Ta-Sn minerals occurring in pegmatites from the area as part of the on-going studies of the pegmatites and granitoids by the authors.

Ayres and Cerny (1982) subdivided the Yellowknife field into three distinct regions: the Prosperous Lake-Prelude Lake area (Northwestern), the Ross Lake-Redout Lake area (Northeastern) and the Southeastern area. In each region, there are geographically related populations of pegmatites (Fig. 1). Series designation used herein is based primarily on geographical distribution, and it has no genetic connotation at present. Names of pegmatites and swarms are based on historical usage where common or names by the authors.

Rare-metal pegmatites from the Yellowknife area are of the beryl + columbite, complex (Li, Be, Nb, Ta) and spodumene types (Cerny and Ercit, 1985). Although the number of granitic pegmatites that occupy the Yellowknife region is overwhelming, only a small fraction of these contain Nb-Ta-Sn minerals, which occur as rare accessories in all types. They are black in colour and range in size from microscopic to 5 cm across. Crystals may be thinly bladed and platy (locally in radiating aggregates), blocky or tabular (Fig. 2). They are found in various paragenetic associations and in all zones of the pegmatite, but occur most frequently in a medium-grained cleavelandite + quartz $\pm$ lithium muscovite assemblage, in fine-grained saccharoidal albite or in replacement bodies of lithium muscovite within the intermediate zone or core of the pegmatite. 534 electron microprobe analyses of 345 Nb-Ta-Sn minerals from 108 rare-metal pegmatites were collected in an attempt to characterize the general geochemistry and fractionation trends of the Yellowknife field (for a review of fractionation trends in the Nb-Ta and Fe-Mn system, the reader is referred to Cerny and others, 1985).

The list of Nb-Ta-Sn minerals found in the Yellowknife region includes members of the columbite-tantalite series (Fe,Mn)(Nb,Ta)₂O₆, ixiolite (Ta,Nb,Fe,Mn,Sn)O₃, wodginite (Mn,Fe)₄(Sn,Ta,Ti)₄(Ta,Nb)₂O₁₂, ferrotapiolite FeTa₂O₆, cassiterite SnO₂ and microlite (Na,Ca)₂Ta₂O₆(OH,F). In addition, an unknown Ca-Nb oxide mineral has been found, but identification has yet to be completed. Members of the columbite-tantalite group are the most abundant species in the field and have been identified in nearly all of the rare-metal pegmatites examined. Ferrocolumbite, manganocolumbite, ferrotantalite and manganotantalite are distinguished on the basis of the atomic ratios in the two cation groups of the general formula AB₂O₆, where A = Fe, Mn and B = Nb, Ta. Ferrotapiolite, cassiterite and microlite are less common, but may be

abundant locally. Ixiolite and wodginite are extremely rare. Representative microprobe analyses of Nb-Ta-Sn oxide minerals from the Yellowknife pegmatite field are provided in Table 1.

SOUTHEASTERN REGION

This region (Fig. 3) is in the extreme southeastern corner of the Hearne Lake (85J) map area. Meintzer and Cerny (1983) provided an initial description of the pegmatite series of this area based on field examination and previous works. In addition they provided sketch maps indicating the geographical location of the various dikes. Nb-Ta-Sn minerals from this area were first reported by Jolliffe (1944).

Faulkner Lake Series

This series surrounds the Faulkner tonalite and contains the most fractionated pegmatites in the southeastern section. The BET, BIG HILL, MOOSE bodies and TAN swarm make up this series. Nb-Ta-Sn minerals are fairly extensive throughout the series, although ferrotapiolite-cassiterite intergrowths were the only Nb-Ta-Sn minerals found in the BIG HILL pegmatite.

Ferrotapiolite is the dominant Nb-Ta mineral present in the TAN swarm. It occurs as short prismatic crystals or anhedral grains disseminated in replacement muscovite + K-feldspar, or as platy crystals in cleavelandite. In addition, cassiterite, ferro- and manganotantalite are found as intergrowths within ferrotapiolite.

In the BET pegmatite Nb-Ta minerals are closely associated with metasomatic albitization. Ferrocolumbite and rarely ferrotantalite occur as blocky or platy crystals in cleavelandite. Rare inclusions of ferrotapiolite or cassiterite occur in ferrotantalite.

Nb-Ta in the MOOSE pegmatite is essentially restricted to ferrocolumbite with one sample containing rare inclusions of cassiterite. The ferrocolumbite occurs as platy, tabular or blocky crystals in cleavelandite $\pm$ lithium muscovite $\pm$ quartz, and rarely with amblygonite. No variation in chemistry is observed around the different morphological or paragenetic types.

Buckham Lake West Series

West of Buckham Lake, 4-13 km from the Buckham Granite, lie the HID and MAC pegmatites, and the LIT swarm. No Nb or Ta-bearing minerals were found in the HID dike and the LIT swarm has a limited content of Nb-Ta minerals; platy to blocky crystals of ferrocolumbite occur in cleavelandite $\pm$ quartz $\pm$ muscovite.

The MAC pegmatite contains several species of Nb-Ta-bearing minerals, and in this respect it is one of the most diverse bodies in the Yellowknife field. Nb-Ta minerals are hosted primarily by pink, medium-grained cleavelandite and are dominated by platy to

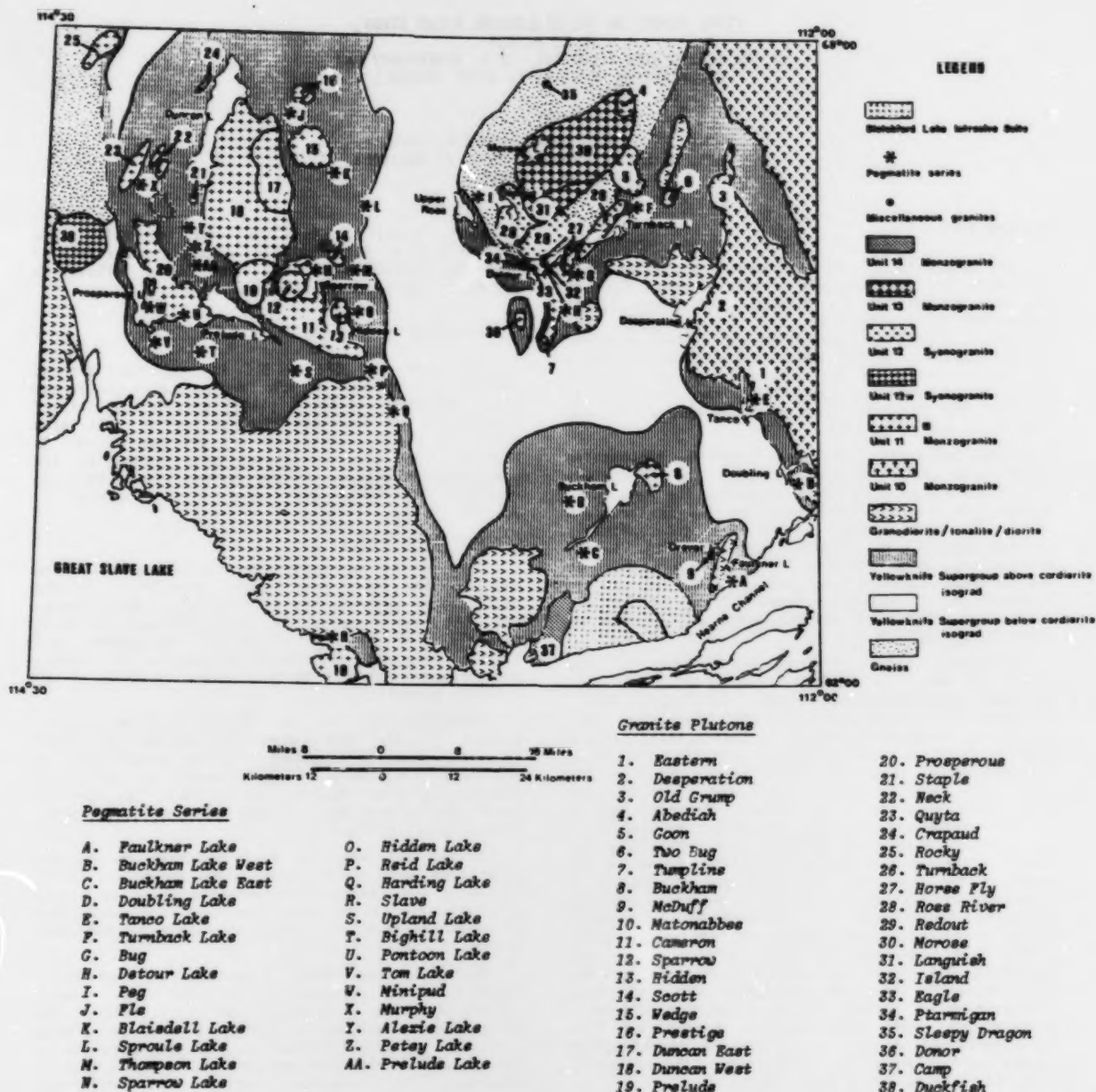


FIGURE 1: The Yellowknife pegmatite field, District of Mackenzie.

tabular ferrotantalite-ferrocolumbite. Ferrotapolite and cassiterite occur rarely as inclusions in ferrotantalite or ixiolite. Manganian- and ferroantantalite ixiolite are common and occur as anhedral massive grains up to 5 cm in diameter. The first occurrence of wadginite from the Yellowknife field is found here; wadginite is intergrown with cassiterite and ixiolite.

Buckham Lake East Series

The pegmatites of this series (MUT, BIN, LENS) lie east of Buckham Lake approximately 8 km southwest of the Buckham Granite. No Nb-Ta-Sn minerals were noted in the MUT and LENS dikes. Only cassiterite was found in the BIN swarm.



FIGURE 2: a) Bladed and radiating aggregates of columbite-tantalite from the PEG Series; b) Single crystals of Nb-Ta oxide minerals. Left: ferrocolumbite from the MOOSE pegmatite; center and right: tapiolites from the PEG Series.



FIGURE 3: Southeastern Region of the Yellowknife Pegmatite Field: Tanco Lake Series: 1 - JO swarm, 2 - THOR-ECHO swarm; Doubling Lake Series: 3 - WAS, 4 - USK; Buchanan Lake West Series: 5 - LIT swarm, 6 - HID, 7 - MAC; Buchanan Lake East Series: 8 - MUT, 9 - BIN swarm, 10 - LENS; and Paulkner Lake Series: 11 - BET, 12 - TAN swarm, 13 - MOOSE, 14 - BIG HILL. Lithologies are the same as Figure 1, except for metamafites near Doubling Lake (wavy pattern) and undivided metasedimentary rocks (no pattern). (Base geology after Henderson, 1976).

Tanco Lake Series

Although this series (JO, THOR-ECHO swarm) shows extensive mineralogical, textural and structural diversity, ferrocolumbite, ferrotantalite and rare manganocolumbite are the only Nb-Ta minerals present. They occur as platy, bladed, tabular, blocky

or anhedral grains within cleavelandite + quartz, K-feldspar + plagioclase + quartz, saccharoidal albite, or rarely, with spodumene or amblygonite. Ferro-tapiolite is rare and occurs intergrown with ferrotantalite in the Peanut dike of the THOR-ECHO swarm (cf. dike PEA in Fig. XI-14 of McIntzer and Cerny, 1983).

Doubling Lake Series

This series consists of the USK and WAS pegmatites situated about 1 km west of the Doubling Granodiorite. Extensive albitization in the form of saccharoidal albite and fine-grained muscovite is common in the USK pegmatite and hosts the majority of the Nb-Ta minerals. In addition, Nb-Ta minerals are also associated with cleavelandite + quartz. Ferrocolumbite is present as blades in both pegmatites and as platy crystals in the WAS. One twinned crystal of ferrotapiolite was found in the USK pegmatite, associated with spodumene, beryl and blue apatite.

Geochemistry of the Southeastern Region

Ferrocolumbite, grading into ferrotantalite, is the dominant mineral species present in this region. Analyses of Nb-Ta-Sn minerals are represented in the $\text{FeNb}_2\text{O}_6\text{-MnNb}_2\text{O}_6\text{-FeTa}_2\text{O}_6\text{-MnTa}_2\text{O}_6$ quadrilateral as illustrated in Fig. 4. The columbite-tantalites show high to intermediate Fe/Mn ratios and highly variable Nb/Ta ratios over the region. The overall trend of the whole region indicates a moderately decreasing Fe/Mn ratio with decreasing Nb/Ta fractionation. However, within individual series a very restricted range of the Fe/Mn ratio, as opposed to the broader span of Nb/Ta ratios, can be seen. The MOOSE pegmatite, for example, contains columbites which belong to three separate morphological and paragenetic types, yet they are chemically uniform throughout the pegmatite. Unlike the general trend for the series, they show very narrow ranges of both Fe/Mn and Nb/Ta ratios.

Locally, the general area shows only minor Sn-enrichment as evidenced by the rare occurrence of

TABLE 1: REPRESENTATIVE MICROPROBE ANALYSES OF Nb-Ta-Sn OXIDE MINERALS FROM THE YELLOWKNIFE PEGMATITE FIELD.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Ta ₂ O ₅	25.5	47.1	47.4	23.1	61.5	63.1	65.8	62.8	81.4	70.8	0.7	13.5	68.0	65.8
Nb ₂ O ₅	55.1	34.7	36.7	57.9	23.1	20.1	15.9	9.8	1.2	11.1	0.3	0.9	7.0	2.9
SnO ₂	0.3	0.4	0.0	0.0	0.0	0.6	3.4	15.0	0.7	0.3	98.7	84.4	0.0	0.0
TiO ₂	0.5	0.3	0.3	0.6	0.3	0.6	0.4	0.5	0.0	1.3	0.0	0.0	0.5	0.3
UO ₂	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.1	12.6
Fe ₂ O ₃ *	0.5	0.7	0.0	1.0	0.0	0.8	0.0	0.0	3.1	1.6	0.4	0.3	0.0	0.0
PbO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	2.7
FeO	15.0	11.7	8.3	9.2	10.6	2.3	6.5	5.8	11.4	12.9	0.0	2.2	0.0	1.6
MnO	3.8	4.9	9.1	9.6	5.5	12.9	7.3	5.3	0.7	0.9	0.0	0.2	0.0	0.0
CaO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	17.2	6.4
Na ₂ O	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.2	1.6
Totals	100.7	99.8	101.8	101.4	101.0	100.4	99.3	99.2	98.5	98.8	100.1	101.4	96.3	93.9
Ta ⁵⁺	0.86	1.77	1.74	0.76	2.45	2.55	0.92	7.20	3.74	3.04	0.01	0.18	3.36	3.68
Nb ⁵⁺	3.09	2.17	2.24	3.17	1.53	1.35	0.37	1.90	0.09	0.79	0.01	0.02	0.57	0.27
Sn ⁴⁺	0.02	0.02	0.00	0.00	0.00	0.04	0.07	2.53	0.05	0.02	1.97	1.69	0.00	0.00
Ti ⁴⁺	0.04	0.03	0.03	0.05	0.03	0.07	0.02	0.16	0.00	0.15	0.00	0.00	0.07	0.05
U ⁴⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.04	0.58
Fe ³⁺	0.05	0.08	0.00	0.09	0.00	0.09	0.00	0.00	0.41	0.18	0.02	0.01	0.00	0.00
Pb ²⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.15
Fe ²⁺	1.55	1.36	0.94	0.93	1.30	0.29	0.28	2.05	1.62	1.70	0.00	0.09	0.00	0.27
Mn ²⁺	0.39	0.57	1.05	0.99	0.69	1.62	0.32	1.89	0.10	0.12	0.00	0.01	0.00	0.00
Ca ²⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	3.34	1.40
Na ⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.78	0.63
Totals	6.00	6.00	6.00	5.99	6.00	6.01	1.98	15.73	6.01	6.00	2.01	2.00	8.17	7.03

1 - ferrocolumbite; 2 - ferrocolumbite; 3 - manganocolumbite; 4 - Fe-rich manganocolumbite; 5 - ferrotantalite; 6 - manganotantalite; 7 - iziolite; 8 - wodginite; 9 - ferrotapiolite; 10 - Nb-rich ferrotapiolite; 11 - cassiterite; 12 - tantalian cassiterite; 13 - microlite; 14 - uranmicrolite. *Fe₂O₃ calculated by stoichiometry. Cations normalized on: 12 oxygens - columbite-tantalite, ferrotapiolite, 32 oxygens - wodginite, 2 oxygens - iziolite, cassiterite, 4(Ta + Nb + Ti) - microlite.

cassiterite. The cassiterite itself is commonly tantalian (up to 14 wt. % Ta₂O₅). Some Sn enrichment occurs in the Faulkner Lake Series in ferrotapiolites and tantalites. Wodginite from the Buckham Lake Series has a low Nb/Ta ratio and intermediate Fe/Mn ratio along with a high Sn content (17 wt. % SnO₂).

NORTHEASTERN REGION

The Northeastern Region dikes include the Turnback Lake, Bug, Detour Lake and Peg Series as delineated by Meintzer and others (1984). Of these, only the Peg Series contain Nb-Ta minerals.

Peg Series

The PEG Series (Fig. 6) is a regionally zoned pegmatite swarm that lies between Upper Ross and Redout Lakes (Hutchinson, 1955). Nb-Ta-Sn minerals in the form of ferrocolumbite, ferrotantalite, manganotantalite, ferrotapiolite, cassiterite and microlite are found predominantly in Hutchinson's zones 3-5 as modified by the authors.

The crystal morphology of columbite-tantalite varies greatly within the series, ranging from platy, bladed and tabular crystals to radiating blades or anhedral grains. Nb-Ta-Sn minerals occur in medium-grained plagioclase (cleavelandite) + quartz + muscovite assemblages. Manganocolumbite occurs only in the spodumene-bearing dikes of zone 5.

Ferrotapiolite is generally found as single, anhedral grains or as blebs within ferrotantalite in

zone 4 of the region. It is found in similar paragenetic associations as the columbite-tantalite and rarely occurs with beryl or biotite.

Cassiterite is present only in the spodumene-bearing pegmatites as subhedral crystals in plagioclase + quartz. Microlite was found in one dike (#93) where it occurs as rare anhedral blebs replacing ferrotapiolite.

Geochemistry of the Northeastern Section

Analyses of columbite-tantalite indicate an increasing Mn/decreasing Fe trend westward from zone 3 to 5; however, there appears to be no obvious trend in Nb-Ta (Fig. 5). The area shows a slight enrichment in Ti as most of the Nb-Ta minerals present have elevated Ti contents. As with Fe, Ti also shows a decreasing trend from zone 3 to 5. Ferrotapiolite, which occurs primarily in zone 4, contains an average 1.5 wt. % (max. 5%) TiO₂. Cassiterites (zone 5) are also enriched in Ti (0.32 wt. % TiO₂) and contain minor Ta, Nb and Fe. The rare microlite replacing ferrotantalite is Ca and Ta rich and contains little or no Ti, U or Pb.

NORTHWESTERN REGION

The Northwestern Region refers to an extensive array of pegmatite dykes situated around the Wedge, Cameron, Sparrow, Hidden, Duncan and Prosperous Granites. The dikes include the pegmatite series (J) to (AA) as listed in Fig. 1. Only those series described below contain Nb-Ta-Sn minerals.

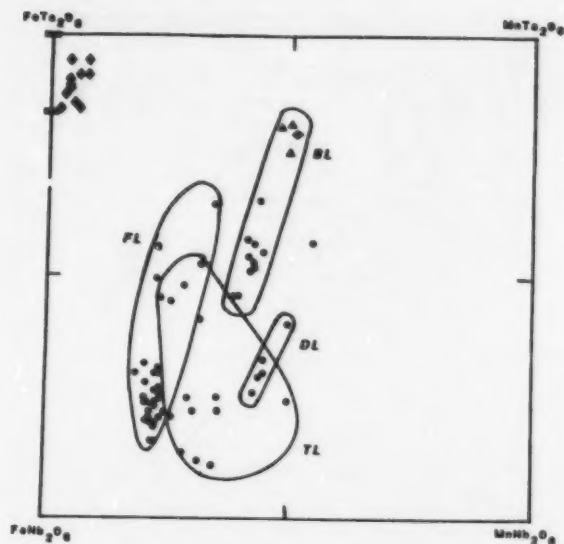


FIGURE 4: Nb-Ta-Sn oxide minerals of the Southeastern Region in the FeNb_2O_6 - MnNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols: (●) - columbite-tantalite, (◐) - wadginitite, (◑) - tapiolite, (■) - cassiterite. FL - Faulkner Lake Series, DL - Doubling Lake Series, TL - Tanco Lake Series.

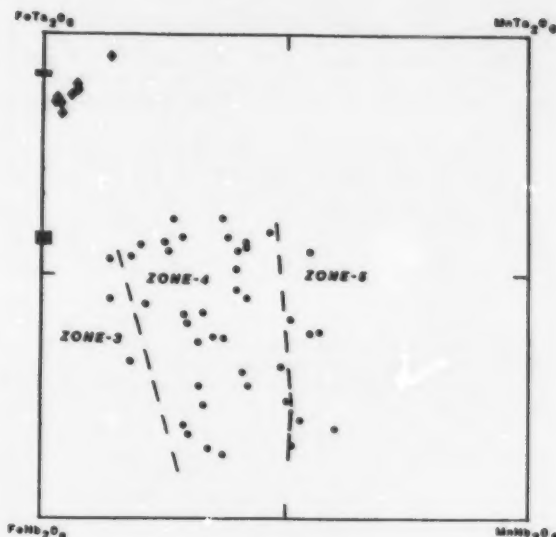


FIGURE 5: Nb-Ta-Sn oxide minerals of the PBG Series in the FeNb_2O_6 - MnNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols as in Fig. 4.

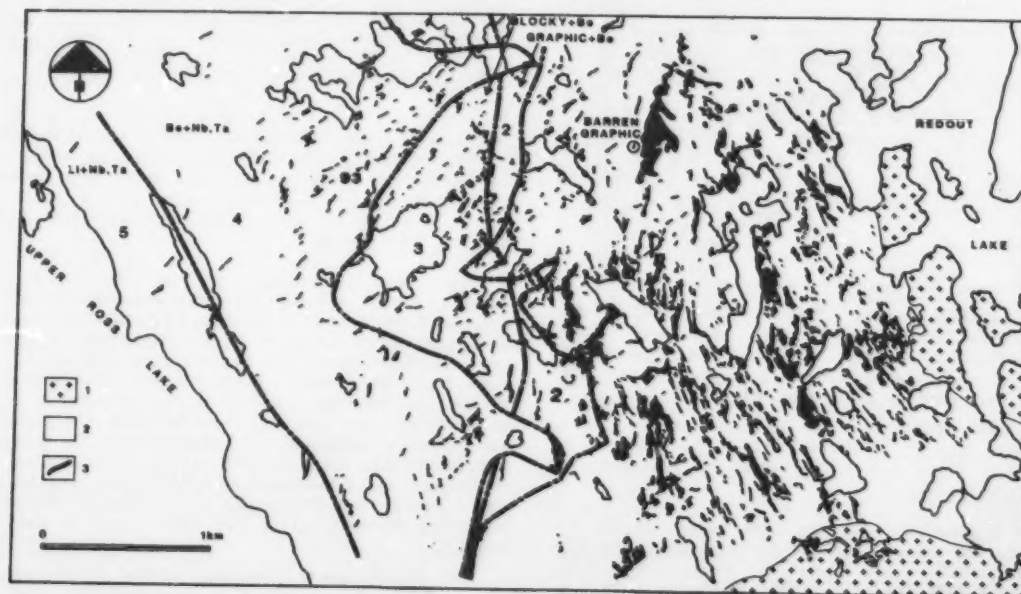


FIGURE 6: Regional zonation of the PBG Series near Redout and Upper Rose Lakes. Symbols: (1) - Redout Granite; (2) - undifferentiated gneisses, metavolcanics and schists; (3) - Pegmatites; ○ - Beryl occurrences in zone 1; ◇ - Nb, Ta oxide occurrences in zone 1. Zonation modified after Hutchinson, 1955.

Blaisdell Lake Series

The Blaisdell Lake Series lies within 3 km east and southeast of the Wedge Granite near Blaisdell Lake and consists of the VO, BILL and MELODY swarms (Fig. 7).

Columbite-tantalite, ferrotapiolite, cassiterite and microlite have been identified in the VO swarm in association with cleavelandite or saccharoidal albite. Columbite-tantalite is present as radiating blades and tabular crystals and microlite as anhedral grains replacing manganotantalite. Cassiterite, however, occurs as microscopic inclusions within ferrotapiolite.

Ferrocolumbite is the only Nb-Ta mineral present in the BILL swarm and occurs as bladed, platy or tabular crystals in plagioclase, K-feldspar $\pm$ quartz, or beryl. Rarely, cassiterite is associated with ferrocolumbite.

The MELODY swarm contains ferrocolumbite, ferrotantalite and ferrotapiolite. Ferrotapiolite occurs as large grains in K-feldspar near core margins, and ferrocolumbite-ferrotantalite forms platy crystals and granular masses associated with cleavelandite or K-feldspar $\pm$ beryl.

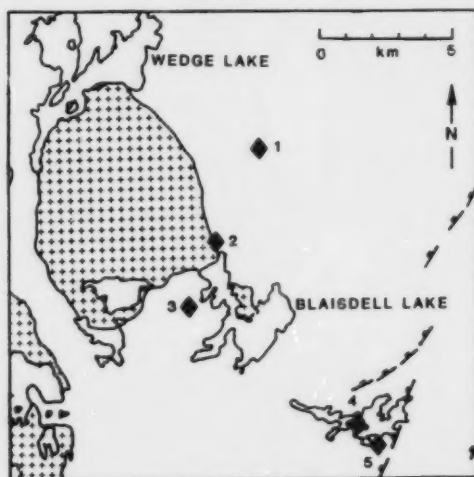


FIGURE 7: Blaisdell Lake - Sproule Lake area of the Yellowknife Pegmatite Field: Blaisdell Lake Series: 1 - VO Swarm, 2 - BILL Swarm, 3 - MELODY Swarm; Sproule Lake Series: 4 - FLY, 5 - CATA Swarm (Base geology after Henderson, 1976).

Sproule Lake Series

The FLY pegmatite and the CATA swarm outcrop in nodular biotite schist south of Sproule Lake. The FLY pegmatite contains thin blades of ferrocolumbite in a quartz + K-feldspar assemblage. Microscopic blades of manganocolumbite also occur in fine-grained albite bodies. Several of these blades show alteration of their rims to fermsite(?).

The CATA swarm is generally rich in cassiterite, which becomes more common in the southernmost extension of the swarm. Cassiterite occurs chiefly in cleavelandite, along with blue-green apatite and

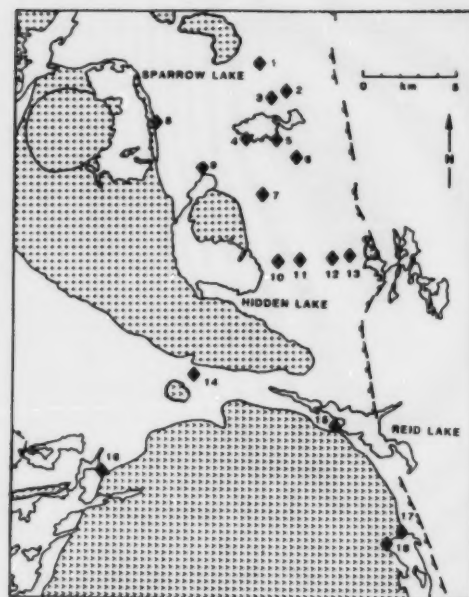


FIGURE 8: The Sparrow - Thompson - Hidden Lakes area of the Yellowknife Pegmatite Field: Thompson Lake Series: 1 - FRED, 2 - TANYA Swarm, 3 - WACO Swarm, 4 - MURKY, 5 - TOM, 6 - KI Swarm, 7 - PI Swarm; Sparrow Lake Series: 8 - CASPER Swarm, 9 - HEIDI; Hidden Lake Series: 10 - GREG Swarm, 11 - QI Swarm, 12 - JIM-LIT Swarm, 13 - LU Swarm; Reid Lake Series: 14 - PANCHO VILLA Swarm, 15 - ANN/BOA Swarm; Upland Lake Series: 16 - JENNE Swarm; Harding Lake Series: 17 - JAKE-DA Swarm, 18 - PAINT (Base geology after Henderson, 1976).

amblygonite. Manganocolumbite is the dominant member of the columbite-tantalite series present, occurring as blades or tabular crystals in cleavelandite $\pm$ muscovite.

Sparrow Lake Series

This series consists of the CASPER and HEIDI swarms, which intrude the eastern margin of the Sparrow/Cameron Granite and surrounding schist, east of Sparrow Lake and north of Hidden Lake (Fig. 8). In the CASPER swarm, Nb-Ta minerals exist in the form of ferrocolumbite-ferrotantalite and rare ferrotapiolite and microlite. Columbite-tantalite is present as anhedral grains and veinlets in muscovite + perthite or plagioclase + quartz $\pm$ muscovite assemblages. Ferrotapiolite occurs as inclusions in ferrotantalite from only one dike in the swarm. Additionally, secondary microlite replacing ferrotapiolite has been found in only one sample.

The HEIDI swarm contains bladed ferrotantalite in K-feldspar + quartz assemblages and in rare, late fine-grained muscovite replacing K-feldspar.

Thompson Lake Series

This series includes a group of pegmatites that extends 5 km north, south and west of Thompson Lake.

Kretz (1968) described the pegmatites of this area extensively, and noted a regional mineralogical zonation in them; rare-element content increasing to the northeast. Nb-Ta-Sn minerals include ferrocolumbite, cassiterite, ferrotapiolite and microlite.

The MURKY and TOM pegmatites are single bodies that lie on the western and southern shores of Thompson Lake. The MURKY body is partly zoned with beryl and columbite present. Ferrocolumbite is found in small veinlets (2.5 x 0.6 cm) in a plagioclase + quartz + muscovite intermediate zone and also as platy crystals in blocky K-feldspar. On the other hand, the TOM pegmatite shows no internal zonation except for alternating bands of aplitic and coarse-grained perthitic pegmatite. Platy crystals of ferrocolumbite are present in quartz-muscovite patches.

North of Thompson Lake lie the WACO swarm, TANYA swarm and FREDA dike in order of increasing distance. Although the authors found no Nb-Ta-Sn minerals in the TANYA swarm, assays of chip samples from these dikes do show elevated Ta_2O_5 contents (Hood, pers. comm.). In the WACO pegmatites, ferrocolumbite is found as platy crystals or anhedral grains in blocky K-feldspar or cleavelandite. Ferrotapiolite and cassiterite form anhedral grains within the ferrocolumbite.

The FREDA pegmatite is the most fractionated body in the area and contains ferrocolumbite, cassiterite, ferrotapiolite and microlite. Ferrocolumbite is associated with cassiterite, beryl, amblygonite or lithium muscovite in K-feldspar or cleavelandite + quartz assemblages. The ferrocolumbite is commonly intergrown with massive cassiterite within patches of replacement muscovite concentrated in the western end of the pegmatite. One sample of ferrotapiolite was found as anhedral blebs within cassiterite. The ferrotapiolite is partly replaced by Ca-rich microlite.

Further to the south lie the FI and KI swarms of which only the former contains Nb-Ta minerals. Nb-Ta minerals are rare in the spodumene-bearing FI pegmatites and occur as bladed crystals of ferrocolumbite in isolated K-feldspar + quartz pods. No Nb-Ta minerals were found in the KI swarm.

Hidden Lake Series

The spodumene-bearing QI, GREG, LU and JIM-LIT pegmatite swarms make up the Hidden Lake Series extending approximately 6 km east of the Hidden Granite. Ferrocolumbite is the dominant Nb-Ta species found in the QI, GREG, and LU pegmatites along with minor ferrotantalite. Ferrocolumbite forms anhedral grains (QI) or bladed crystals (GREG, LU) in cleavelandite + quartz + muscovite $\pm$ spodumene. Cassiterite is abundant only in the LU and JIM-LIT pegmatites where it occurs as blocky to subhedral crystals in plagioclase $\pm$ quartz $\pm$ muscovite.

Reid Lake Series

This series comprises the ANN dike, BOA swarm and PANCHO VILLA swarm pegmatites. All pegmatites contain abundant spodumene and outcrop within 2 km of the southeastern margin of the Cameron Granite. Nb-Ta minerals are present as fine-grained blades of ferrocolumbite and ferrotantalite in a cleavelandite + quartz + muscovite assemblage in the ANN dike and PANCHO VILLA swarm along with rare

grains of ferrotapiolite in the PANCHO VILLA swarm. No Nb-Ta-Sn minerals have been found in the BOA swarm.

Harding Lake Series

The JAKE-DA and PAINT pegmatites are north and northwest of Harding Lake and intrude the margins of the Southeastern Granodiorite. Manganocolumbite is the only Nb-Ta-bearing mineral present. It forms thin blades in replacement muscovite or radiating plates or blocky crystals in a cleavelandite + quartz + muscovite assemblage.

Upland Lake Series

The PANCHO PETE and JENNE pegmatites make up the Upland Lake series, however only the JENNE dikes contain Nb-Ta minerals. The spodumene-bearing JENNE dikes are homogeneous with tiny blades of ferrocolumbite present in plagioclase + muscovite + quartz.

Bighill Lake Series

The LIMO and BIG swarms, situated respectively on the northern and southeastern shores of Bighill Lake, make up this series (Fig. 9). The LIMO dikes contain only rare platy crystals of ferrocolumbite in saccharoidal albite.

In the BIG pegmatites, ferrocolumbite and manganocolumbite are the dominant Nb-Ta minerals and occur as platy to tabular crystals in plagioclase + quartz $\pm$ muscovite. Ferrotapiolite and cassiterite form inclusions within columbite-tantalite. In addition, rare uranian microlite replaces ferrotapiolite.



FIGURE 9. Bighill Lake area of the Yellowknife Pegmatite Field: Tom Lake Series: 1 - ODIN Swarm, 2 - NITE Swarm, 3 - MINT Swarm; Bighill Lake Series: 4 - LIMO Swarm, 5 - BIG Swarm (Base geology after Henderson, 1976).

Tom Lake Series

This series (ODIN, NITE and MINT swarms) is approximately 3 km south of the Prosperous Granite and north and west of Bighill Lake. Platy crystals of manganocolumbite are formed in the ODIN pegmatite. Blocky to subhedral, tabular crystals of manganotantalite are common in the MINT and NITE pegmatites in cleavelandite + muscovite and rarely with spodumene and quartz. Tantalian ixiolite with

Mn = Fe, intergrown with manganotantalite and minor ferrotapioilite, is also present in the MINT and NITE pegmatites.

Petey Lake Series

Two major pegmatite swarms, the Lil' Rascals and the Geo swarms, penetrate schist of the Burwash Formation between the Prosperous and Duncan Granites. Only the DR. BOB dike of the Geo swarm and the RIBER pegmatite contain Nb-Ta-Sn minerals (Fig. 10). In the DR. BOB dike, ferrocolumbite forms inclusions within anhedral grains of cassiterite associated with medium-grained cleavelandite.

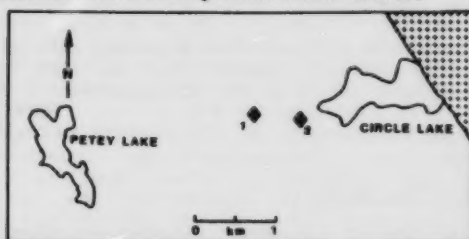


FIGURE 10: Location of the DR. BOB and RIBER pegmatites (Petey Lake Series) of the Yellowknife Pegmatite Field (approx. 5 km SSE of Alezie Lake). 1 - DR. BOB, 2 - RIBER.

The RIBER pegmatite is the most fractionated and mineralogically diverse pegmatite in the series. Nb-Ta minerals are represented by blocky to prismatic manganotantalite along with subordinate platy ferrocolumbite in cleavelandite $\pm$ quartz $\pm$ muscovite $\pm$ rare amblygonite and tourmaline. Uranian microlite with variable U, Pb and Ca concentrations occurs as isolated subhedral grains or anhedral blebs replacing manganotantalite.

Geochemistry of the Northwestern Region

Columbite-tantalites from the Blaisdell Lake Series show the limited Fe/Mn ratios and variable Nb/Ta ratios characteristic of the Yellowknife field (Fig. 11). Fe-Mn fractionation suggests a regional zonation about the Wedge Granite, with the least fractionated (highest Nb/Ta ratio) BILL swarm lying near the margins of the granite-schist contact, the more fractionated MELODY pegmatites further out and the most fractionated, spodumene-bearing VO swarm at the greatest distance from the Wedge Granite (lowest Nb/Ta ratio).

Field observations from the Thompson, Sparrow and Hidden Lakes series indicate the presence of another regionally zoned pegmatite area around the Sparrow/Cameron and Hidden biotite-muscovite granites. As in the other series, a restricted Fe/Mn ratio is manifested by a steep decrease in the Nb/Ta ratio with increasing fractionation (Fig. 12). From field observations and laboratory data, it appears that pegmatite fractionation increases eastward from the Sparrow/Cameron Granite. High Nb content and extreme Mn fractionation of the JAKE-DA and PAINT bodies suggest that these may not be genetically related to the pegmatites of the Sparrow, Thompson, Hidden and Reid Lakes Series.

The Bighill Lake and Tom Lake Series show relatively high Ta and Mn fractionation and may represent the peak of pegmatite fractionation in a

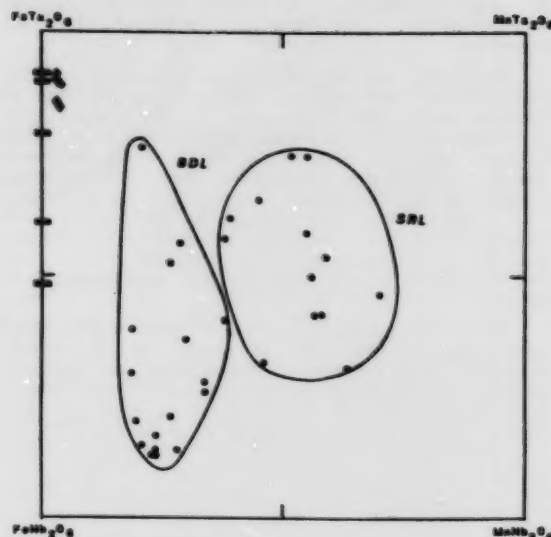


FIGURE 11: Nb-Ta-Sn oxide minerals of the Blaisdell - Sproule Lakes area in the FeNb_2O_6 - NbNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols as in Fig. 4. BDL - Blaisdell Lake Series, SPL - Sproule Lake Series.

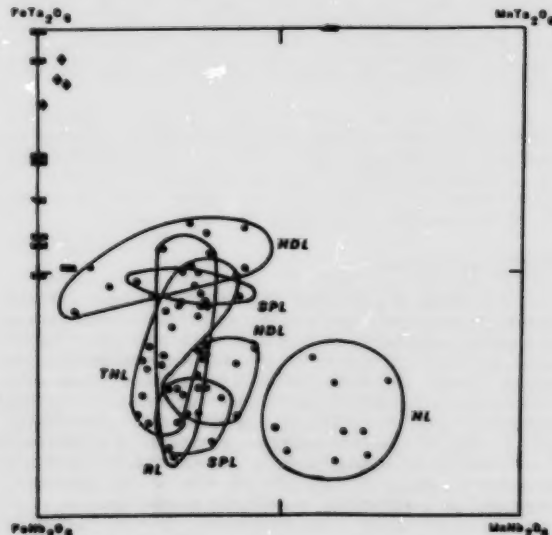


FIGURE 12: Nb-Ta-Sn oxide minerals of the Thompson-Sparrow-Hidden Lakes area in the FeNb_2O_6 - NbNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols as in Fig. 4. SPL - Sparrow Lake Series, HDL - Hidden Lake Series, THL - Thompson Lake Series, RL - Reid Lake Series.

southwesterly direction from the Prosperous two-mica granite (Fig. 13).

Although Nb-Ta minerals are rare in the pegmatites between the Prosperous and Duncan Granites, the area contains the RIBER pegmatite, the most fractionated dike in the region. Nb-Ta and Fe-Mn fractionation in this body proceeds normally (decreasing Nb/Ta, decreasing Fe/Mn), with a reversal in the Nb-Ta fractionation in later stages of pegmatite crystallization. This reversal is probably caused by the crystallization of microlite into which Ta is partitioned preferentially to tantalite.

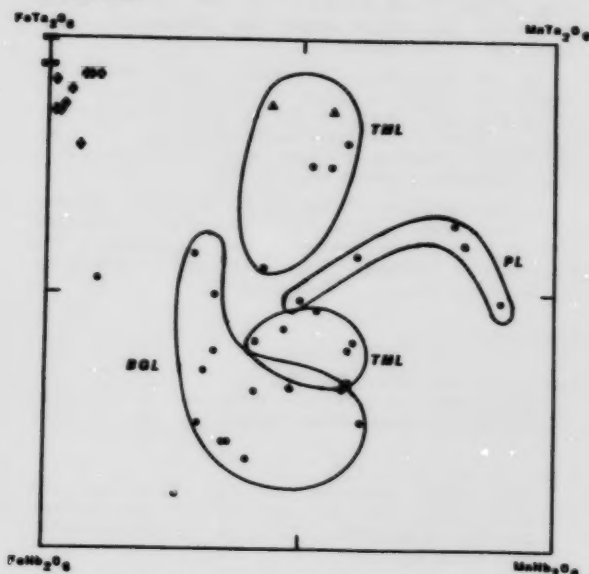


FIGURE 13: Nb-Ta-Sn oxide minerals of the Bighill and Circle Lakes areas in the FeNb_2O_6 - MnNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols as in Fig. 4. PL - Patsy Lake Series, TNL - Tom Lake Series, BGL - Bighill Lake Series.

SUMMARY

Chemistry of Ferrotapiolite, Cassiterite and Microlite

The chemistry of ferrotapiolites from the Yellowknife area is rarely variable. Mn contents are generally low, as are Ti and Sn. Coexisting ferrotapiolite-tantalite pairs indicate the preferred partitioning of Fe and Ta into the tapiolite while Mn and Nb prefer the tantalite (Fig. 14).

Cassiterite follows a similar partitioning effect and, as with ferrotapiolite, contains little or no Mn and Ti (Fig. 15). Fe, Ta and Nb are commonly present in cassiterites.

Microlite from the Yellowknife field is generally high in Ca and Ta and low in Nb, Ti, Sn, U and Pb (Fig. 16). Some microlites contain appreciable U and Pb, with up to 14 wt. % UO_2 and 4 wt. % PbO . Most microlites tend to have higher Fe/Mn and lower Nb/Ta ratios than tantalite and higher Nb/Ta ratios than ferrotapiolite.

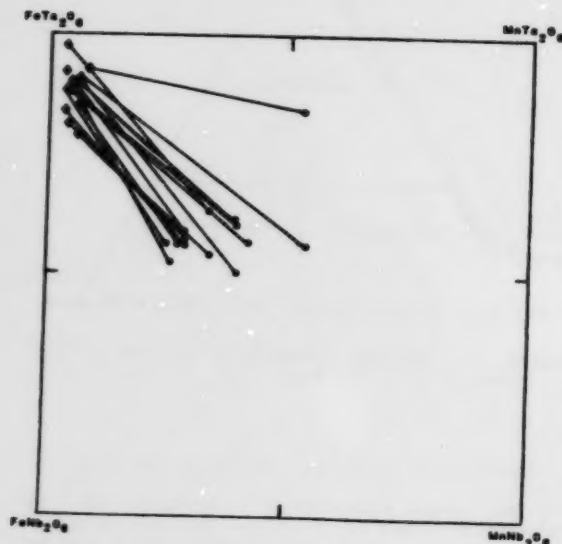


FIGURE 14: Coexisting tapiolite-tantalite pairs in the FeNb_2O_6 - MnNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols as in Fig. 4.

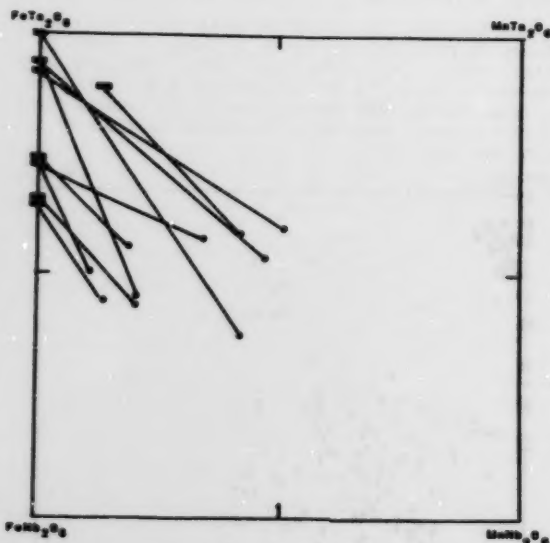


FIGURE 15: Coexisting cassiterite-tantalite pairs in the FeNb_2O_6 - MnNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols as in Fig. 4.

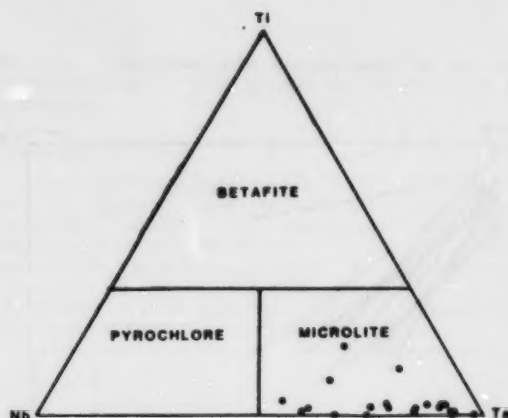


FIGURE 16: Microcline composition in the Ta-Nb-Ti triangular plot (at. %).

Chemistry of Columbite-Tantalite and Ixiolite

The gross geochemical signature of the Yellowknife pegmatite field is that of Fe and Nb dominance, with only local enrichments in Ta, Sn and Ti. This is evident from the clustering of columbite-tantalite analyses in the Fe-Nb quadrant of the quadrilateral (Fig. 17). Mn-rich columbite-tantalites are rare. Most of the pegmatite series generally show highly variable Nb/Ta ratios and restricted Fe/Mn ratios in the columbite-tantalite group.

In contrast to columbite-tantalite, ixiolite is always Ta-rich and the ratio of Fe:Mn is close to 1 (Fig. 17). Its SnO_2 and TiO_2 contents are somewhat variable (1.67 - 16.58 and 0.00 - 0.39 wt. %, respectively). The composition of the ordered forms of this mineral, wadginites, is similar to those of Sn-rich ixiolites.

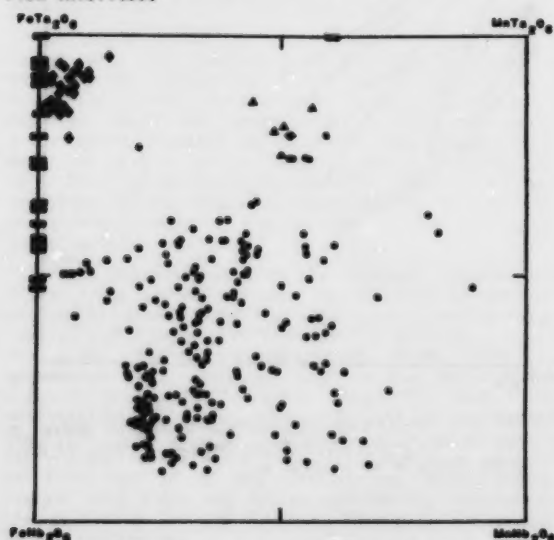


FIGURE 17: Nb-Ta-Sn oxide minerals of the Yellowknife pegmatite field in the FeNb_2O_6 - MnNb_2O_6 - FeTa_2O_6 - MnTa_2O_6 (mol. %) diagram. Symbols as in Fig. 4.

GENERAL CONCLUSIONS

The restricted degree of Fe/Mn fractionation in the Nb and Ta-bearing oxide minerals of the Yellowknife field may be related to the generally low activity of fluorine in the parent pegmatites. As shown by Cerny and Ercit (1985) and Cerny and others (1985), the Fe/Mn ratio rapidly decreases in columbite-tantalites of F-rich pegmatites to the point of virtual elimination of Fe before the onset of noticeable enrichment in Ta. The pegmatites of the Yellowknife field show an opposite tendency: Nb/Ta decreases without appreciable Fe/Mn reduction at any stage of the Ta environment. So far, such a trend appears to be unique among the documented cases (Cerny and Ercit, 1985).

It is noteworthy in this report that the Yellowknife pegmatite field is virtually devoid of lepidolite, and microcline is in general very rare. Early, blocky amblygonite-montebrazite carries substantial fluorine in some pegmatites, but late-stage F-bearing minerals are missing.

Further studies of the Nb-Ta-Sn-bearing species and associated pegmatite minerals, conducted currently by the authors, will correlate the fractionation patterns of the different rock-forming and accessory phases.

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PARAGNEISSES AND GRANITOIDS OF THE SLAVE PROVINCE
CHRISTIE BAY AND RELIANCE MAP SHEETS, BARNSTON RIVER AREA

by

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ABSTRACT

Suites of paragneisses and granitoids from east and west of Barnston River (110°10'W, 62°56'N), south-east Slave Province, have been analyzed for major elements and their Rb-Sr age. The paragneisses are of sillimanite-cordierite grade, higher in the west and lower in the east. They have 60-70% SiO₂ and relatively high MgO, CaO and Al₂O₃. Derived from a turbidite sequence, these gneisses have a very uniform composition, varied only in any significant sense by loss of K during chloritization of biotite. Their discordant Rb-Sr age pattern indicates that this phenomenon is younger than 2540 Ma, the age of the granitic intrusives. Zircons from the paragneiss yield an upper age limit of 2640 Ma, believed to be the provenience age. No older basement has yet been seen in this area.

INTRODUCTION

The regional metamorphism of the southeastern Slave Province increases eastwards (Thompson, 1978). As a consequence of this and because of a dearth of greenstones in the southeastern part of the Province, there are severe problems in devising a stratigraphy or a geological history for this area. Very few radiometric ages have been published.

Mapping at 1:250,000 has shown that much of the region immediately north of the East Arm of Great Slave Lake consists of granitoids (Wright, 1951; map 1123A, Reliance, NTS 75K). Between these, belts of paragneiss or metasediment are preserved, relics of a much larger region of metasediments, with minor quantities of amphibolite. As 3000 Ma rocks are known from Yellowknife (Nikic and others, 1980) and north of Yellowknife (Frith and others, 1974), it was considered that a study of the paragneisses of the accessible Barnston River area (Fig. 1) would assist in defining events in the southeastern part of the Province.

Two reconnaissance visits showed that these paragneisses could be variously described as schists and gneisses in the field. These studies also showed that the amphibolites in the central and southern parts of the Walmsley Lake map sheet (NTS 75N) are not likely to be fragments of a greenstone basement, but are more likely to be local masses of varied volcanics, including silicic pyroclastics. Further, the paragneisses of the same area appear to be Yellowknife Group sediments, metamorphosed locally to high grade by the local granitoids.

The paragneisses of the Barnston River area were studied more extensively in the field and 110 large samples were collected. Half of these have been sectioned, 85 have been analyzed and Rb-Sr and zircon ages obtained. In this report, the rocks are classified by major elements analysis; trace element data will form the subject of a second report. The area studied measures 25 x 5 km, from 109°50' to 110°20' W, and includes units previously mapped as knotted schists, paragneiss, gneissic complex, granite, granodiorite and muscovite granite.

Our conclusion is straightforward: the area formed during the late stages of the history of the Slave Province and there is no evidence for the existence of an earlier basement in this locality.

GEOCHEMISTRY OF THE METASEDIMENTS AND GRANITOIDS

Three main suites of metasediments were collected on traverses across the strike of the three main belts mapped by the GSC (Fig. 1). One, west of Barnston River, is a varied suite of cordierite gneisses; a second similar suite occurs east of Barnston River for a distance of 5 km. A third suite, a further 5 km east, appears to be of lower grade, consisting largely of biotite schist, originally greywacke, but these also contain sillimanite. Close to granitoids, these units become gneissic and intermingled in some cases with granitic sheets. In many areas, it was difficult or impossible to distinguish metasediments from intrusions.

Almost all of these rocks were analyzed by XRF at Durham University. The analyses show that the paragneisses and schistose metagreywackes form a geochemically coherent suite, uniformly characterized by SiO₂ 60-70%, total Fe/MgO around 1.5, variable but low CaO, and Na₂O generally greater than K₂O (Table 1). Each map unit and traverse contained most of the chemical variants. As argued below, some of the variation may be attributed to hydrolysis of biotite, cordierite and plagioclase. In several instances, there is such extensive alteration of modally significant minerals that the tabulated compositions are misleading as indices of the original sedimentary compositions. Further discussion is found below, as it has bearing on the interpretation of the Rb-Sr data, but a full discussion is deferred to a future paper on trace elements. The granitoids are more straightforward, but even some of these have lost some alkalis during chloritization of biotite.

The samples were divided into five chemical types of metasediment, while five classes of granitoids were found. These are characterized in the ensuing section.

GEOCHEMISTRY - DETAILS

Paragneiss 1 : This chemically-defined unit occurs east and west of Barnston River: it is defined by normative anorthite <8% and normative corundum <5%. All five samples are biotite-poor metasediments with comparatively sodic plagioclase, representing an end member of the possible range of variations in bulk composition of the original sediments (Table 1, type

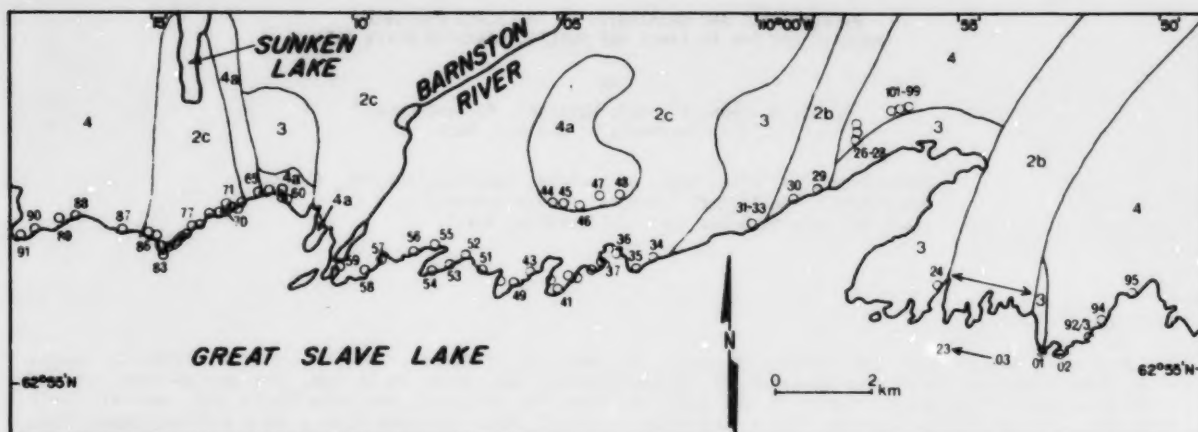


FIGURE 1: Geology and sample locations, Barnston River area. Geology from GSC maps 1122A and 1123A (NTS 75K and 75L). 2b = knotted quartz-biotite schists and hornfels; 2c = paragneiss derived from 2b; 3 = undifferentiated gneisses, including migmatites and mylonites; 4 = granite and granodiorite; 4a = muscovite granite, granodiorite and pegmatite.

1). They did not come from any particular class of metasediment in the field.

Paragneiss 2 : This widespread rock type includes both fine- to medium-grained biotite-plagioclase-quartz metagreywackes (with minor cordierite and sillimanite) and coarser grained cordierite-sillimanite gneisses which are interbedded and may merely represent several psammitic varieties of the original turbidites. Chemically, they all have normative anorthite 8% and normative corundum 5%, features shared by samples with or without highly chloritized biotite (Table 1, type 2). The other main chemical feature, the variable Na/K ratio, is largely a function of the degree of alteration of biotite (as observed in thin section), a phenomenon which is highly variable. Furthermore, in some rocks, muscovite plus chlorite have replaced biotite, while others have no K-mineral pseudomorphs after biotite. As a result, the division into two types (K_2O Na_2O , Na_2O K_2O) is a little on the arbitrary side. However, it is, overall, a reflection of the degree of biotite alteration, as biotite forms, or formed, 20-30% of most of these rocks. Cordierite is always wholly pinitized in this suite; it may, rarely, enclose some anthophyllite. No K-feldspar, primary muscovite or garnet were found. Sillimanite is invariably a replacement for biotite, occurring as swarms of needles or small, highly elongate, prismatic crystals. In only one sample is sillimanite of a grain size approaching that of the other minerals. Tourmaline is a common minor phase, while all the coarser grained cordierite gneisses have abundant zircons as inclusions in biotite. Overall, it is considered that Paragneiss 2 (Na) approaches the primary composition, but 2 (K) is hydrothermally altered.

Paragneiss 3 : The chemical definition of this type is normative corundum 5% and normative anorthite 8% (Table 1, type 3). Thin sections are available for half of these samples: they clearly show that this category includes almost all the samples which have the least chloritization, sericitization, etc., while, conversely, there are no highly altered samples in this group. The only obvious sub-set is one poor in Fe, Mg and related elements. These are poorer in biotite and cordierite than the remainder.

The mineralogy of this group resembles that of Paragneiss 2; the cordierite is invariably pinitized.

Paragneiss 4 : These are the metasedimentary samples which have the most extreme compositions, with low SiO_2 , normative anorthite 8% and normative corundum 5%, an unusual combination (Table 1, type 4). This results from high biotite and oligoclase contents, presumably with a highly aluminous biotite. One sample (85) included in this group does not meet the definition in detail, but it is almost in this category and otherwise resembles this group more than the others.

One sample, called an intermediate paragneiss, seems to be an extreme end member of this series, being very rich in biotite (Table 1, type 5).

Granitoids : This group is very diverse, including samples with very little of one or other feldspar. It includes representatives of the mapped later intrusions and smaller, unmapped granite sheets in the metasediments. Two types of granite (*sensu stricto*) are recognized, with a boundary at $MgO = 1\%$ and a pronounced bimodal distribution of Fe, Mg and Ti (Table 2). The smaller subset, rich in Mg, occurs at either end of the sampled region. Small groups of K-rich and Na-rich granites were sampled, again spread across the area, while three granodiorites from the easternmost granitoid were studied. Their major element chemistry directly reflects the simple mineralogy. Most samples are hololeucocratic quartz-oligoclase-microcline granites with biotite and muscovite in the range 1-5%. Extensive subsolidus recrystallization of the feldspars has occurred. Biotite is characteristically retrogressed to chlorite and muscovite has continued to grow under subsolidus conditions, projecting vermicular or spiky overgrowths into adjacent feldspars. Garnet and tourmaline are rare minor constituents.

The granitoids have an unweighted Fe/MgO of 1.8 and an overall $Fe + MgO$ of 3.3 from Table 2. On a weighted basis, the former rises to 2 and the total falls to 2.5. These features clearly distinguish the granites from all the metasediments except the small low-Fe subset of Paragneiss 3.

TABLE 1: BARNSTON RIVER PARAGNEISS AVERAGES

Type	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	TiO ₂	MnO	S	P ₂ O ₅	n
1)	70.87	14.49	0.95	2.93	2.27	0.99	3.92	2.78	0.55	0.04	0.06	0.17	5
	0.46	0.13	0.13	0.40	0.39	0.17	0.14	0.46	0.08	0.01	0.02	0.01	
	2.58	0.76	0.77	2.40	2.29	0.81	0.82	2.10	0.43	0.04	0.14	0.07	
2A)	63.15	17.96	1.52	4.62	4.12	1.09	2.78	3.69	0.84	0.05	0.06	0.14	14
	0.55	0.25	0.05	0.15	0.17	0.09	0.11	0.12	0.03	0.00	0.01	0.01	
	6.84	3.71	0.79	2.36	2.50	1.03	1.18	1.54	0.44	0.06	0.11	0.10	
2B)	67.62	15.98	1.20	3.79	3.53	1.19	3.49	2.17	0.73	0.05	0.12	0.16	15
	0.35	0.16	0.06	0.12	0.12	0.11	0.12	0.09	0.03	0.00	0.02	0.01	
	4.85	1.61	0.90	1.61	1.64	1.18	1.64	1.40	0.40	0.05	0.36	0.10	
3A)	66.46	15.84	1.20	3.70	3.35	2.43	3.51	2.47	0.71	0.07	0.12	0.20	14
	0.57	0.21	0.07	0.20	0.17	0.15	0.09	0.13	0.02	0.00	0.02	0.01	
	7.13	3.34	0.90	2.77	2.77	1.75	1.29	1.70	0.35	0.06	0.24	0.18	
3B)	71.53	15.90	0.55	1.74	1.06	2.25	4.32	2.18	0.29	0.04	0.03	0.14	4
	0.36	0.48	0.03	0.09	0.16	0.14	0.09	0.24	0.03	0.00	0.02	0.01	
	1.74	1.88	0.14	0.42	0.68	0.67	0.38	1.06	0.13	0.01	0.06	0.05	
4)	63.40	17.08	1.45	4.43	4.22	1.94	3.36	2.94	0.85	0.06	0.16	0.18	6
	1.04	0.29	0.01	0.28	0.28	0.05	0.14	0.21	0.05	0.01	0.04	0.01	
	6.25	1.64	0.66	1.88	1.95	0.34	0.81	1.24	0.40	0.04	0.26	0.05	
5)	56.25	20.31	2.08	6.20	6.07	0.50	1.66	5.52	1.13	0.08	0.07	0.16	1

* The first row is the mean, the second is the standard error and the third row is the range.

However, this subset is cordierite-sillimanite gneiss with high Na₂O/K₂O. Indeed, this chemical criterion can be used with confidence to detect the granitic component in the "gneissic complexes" of the area.

GEOCHRONOLOGY

A suite of samples was analyzed for Rb-Sr at the University of Alberta using standard techniques for isotope dilution and isotope ratio analysis. At the time of analysis, the Micromass 30 yielded a value of 0.71023 for Sr_m 987; ⁸⁷Sr/⁸⁶Sr is quoted uncorrected relative to that value. Reproducibility checks on Rb and Sr were better than ± 1%. Zircon analyses were made using the same procedures and standard and spike solutions as are used by H. Baadsgaard: the apparent ages of the zircons were computed by using a program based on the algorithm of Ludwig (1980). Following Baadsgaard (1983), errors in individual zircon data points are estimated at ± 0.5% in both ratios, a figure slightly higher than is given in the reference. However, the zircon data points scatter slightly more than this error.

The Rb-Sr analyses (Table 3, Fig. 2) were obtained from the late intrusive granites and from a range of metasediments, with varying degrees of alteration. The three most radiogenic granites and the least radiogenic metasediment define an isochron of 2542 ± 4 Ma, IR = 0.7027 ± 0.0001, MSWD = 1.5. The four granites including 48, are not so perfectly co-linear at 2550 ± 35 Ma, IR = 0.7008 ± 0.0012. The three most radiogenic granites and the four least

radiogenic metasediments define a similar line at 2500 ± 13 Ma, IR = 0.7023 ± 0.0002. Combining unrelated rocks on an isochron lacks strict validity, but the results indicate that the granites are probably at least 2540 Ma old, IR = 0.702. This comparatively low initial ratio is close to that of Lambert et al. (1984) for the Kam Formation, Yellowknife Greenstone Belt (0.7017 ± 0.00014) and at the upper limit of the probable range of mantle growth curves for 2600 Ma (0.7011 to 0.7017; Lambert, 1981).

The metasediments did not yield useful chronological information by Rb-Sr. They have clearly been open systems with loss of ⁸⁷Sr long after their formation, the five together giving an isochron of 2270 ± 70 Ma, IR = 0.7049 ± 0.0023. The most altered sample (53) has chloritized biotite and pinitized cordierite, and lies well off the line defined by the other four, which give 2550 ± 130 Ma. It is evident that this alteration was either Proterozoic in age and was a non-equilibrium process at that, or perhaps was more recent and irregular in incidence. That the former is the cause seems more likely, as there is known tectonic disturbance and igneous activity immediately to the south in the lower- to mid-Proterozoic. But the zircons may have been "episodically" re-set about 100 Ma ago, an event which might be reflected by ⁸⁷Sr loss from the fine-grained alteration products in the metasediments.

The zircon results (Table 4, Fig. 3) are from a set separated by grain size, magnetic and hand-picking techniques, taken from a biotite-plagioclase-quartz paragneiss. Poor instrumental sensitivity

TABLE 2: BARNSTON RIVER GRANITOID AVERAGES

Type	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	TiO ₂	MnO	S	P ₂ O ₅	n
1)	71.20	15.12	0.802	2.48	2.10	0.74	5.41	1.42	0.50	0.04	0.00	0.18	2
	1.54	0.54	0.04	0.12	0.80	0.50	0.27	0.14	0.06	0.00	0.00	0.02	
	3.09	1.09	0.07	0.23	1.61	1.01	0.54	0.28	0.11	0.01	0.01	0.05	
2)	72.10	15.17	0.34	1.04	0.64	0.09	2.75	6.52	0.24	0.03	0.00	0.29	5
	0.90	0.40	0.04	0.14	0.09	0.11	0.10	0.58	0.03	0.00	0.00	0.06	
	5.53	2.12	0.24	0.75	0.46	0.61	0.58	3.36	0.14	0.01	0.00	0.35	
3)	73.54	15.47	0.15	0.50	0.17	0.86	4.02	5.06	0.04	0.04	0.00	0.16	9
	0.23	0.12	0.02	0.08	0.03	0.05	0.15	0.27	0.02	0.00	0.00	0.01	
	2.33	1.16	0.26	0.83	0.32	0.50	1.63	2.40	0.17	0.03	0.00	0.06	
4)	69.32	15.37	0.84	2.58	1.90	1.02	4.20	4.03	0.41	0.04	0.01	0.25	5
	0.18	0.28	0.03	0.09	0.15	0.13	0.12	0.09	0.03	0.00	0.00	0.02	
	0.91	1.47	0.16	0.51	0.69	0.75	0.71	0.46	0.15	0.01	0.02	0.01	

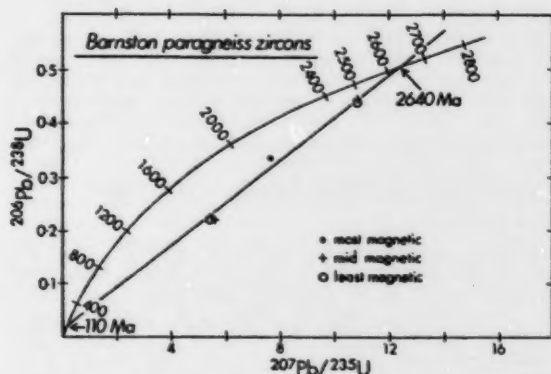


FIGURE 2: Rb-Sr analysis of granite and paragneiss.

prevented use of very small sample techniques, but the results, on aliquots ranging from 4 to 17 mg, show no indication at all of the presence of any significant proportions of pre-2640 Ma zircons. The six aliquots lie along, or very close to, a linear discordant array which intersects concordia at 2640 ± 50 and 110 ± 65 Ma. The former is probably the age (or the mean age of a wider population) of the granitoid or siliceous volcanic sources of the zircons. The little information which we have from the southern Slave Province indicates that this is possible: the Yellowknife Greenstone Belt is 2650 ± 25 Ma old and the Western Granodiorite at Yellowknife is 2635 ± 30 Ma (Lambert et al., 1984 and paper in preparation). The younger intercept may indicate a combination of modern Pb loss or continuous diffusion loss combined with some degree of re-setting during the Proterozoic by ill-defined events, such as those mentioned in the Rb-Sr discussion above. However, the zircons may have been disturbed by large-scale groundwater movements during the maximum of transgression of the Cretaceous sea, which could well

TABLE 3: BARNSTON RIVER AREA Rb-Sr DATA

Sample	Locality	Rb ppm	Sr ppm	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr
Granite					
01	109°53'W 62°55.3'N	111	123	2.633	0.77582 ± 3
32	110°0.2'W 62°56.6'N	130	122	3.114	0.80959 ± 2
37	110°4.0'W 62°56.2'N	255	59 ^x	13.08	1.18389 ± 5
44	110°5.1'W 62°57.0'N	141	87 ^x	4.76	0.87813 ± 5
47	110°4.0'W 62°57.8'N	249	73 ^x	10.25	1.07819 ± 7
48	110°3.5'W 62°57.8'N	276	515	1.557	0.75734 ± 1
Paragneiss					
03	109°53.5'W 62°55.8'N	104	266	1.129	0.74221 ± 3
39	110°4.7'W 62°56.1'N	152	347 ^x	1.273 ^x	0.74962 ± 3
50A	110°6.5'W 62°56.0'N	65	454 ^x	0.413	0.71795 ± 6
					0.71791 ± 5
53	110°7.7'W 62°56.2'N	131 ^x	131 ^x	2.92 ^x	0.79940 ± 4
6640A	109°47'W 63°2.6'N	61	344	0.514	0.71976 ± 3

x = IRP measurement

have reached this area. The regional geology indicates that this part of the surface of the Shield was probably almost at the Cretaceous surface.

The zircon population in sample 6640B consists almost entirely of well-rounded grains of about 0.05 mm diameter, with a few larger, subhedral grains to 0.2 mm length. The latter are elongate but have well-rounded terminations, indicating abrasion during sedimentation. No overgrowths were seen. This gneiss came from a non-migmatitic area, being a representative sample of the thick biotite-plagioclase-quartz gneiss exposed at Misty Lake, 25 km at 60° from the mouth of Barnston River. This particular specimen contains no cordierite, but has a few corroded fragments of garnet.

CONCLUSIONS

The paragneiss belts east and west of the mouth of Barnston River are in a K-feldspar-free, sillimanite-cordierite metamorphic facies. East of

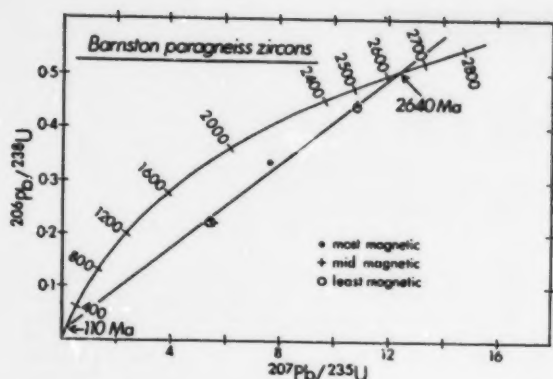


FIGURE 3: Barnston paragneiss zircon analysis.

TABLE 4: BARNSTON RIVER PARAGNEISS ZIRCONS

Aliquot							
Size	Magnetism	^{238}U	^{206}Pb	^{206}Pb	^{207}Pb	^{206}Pb	^{207}Pb
Nicrons		ppm	ppm	^{204}Pb	^{204}Pb	^{236}U	^{235}U
>125	High	542	157.6	150.96	37.43	0.3358	7.674
125-50	High	775	152.6	137.90	36.34	0.2275	5.473
>125	Mid	319	121.6	163.49	41.53	0.4403	10.887
125-50	Mid	344	66.4	177.84	43.71	0.2228	5.435
>125	Low	274	104.2	189.76	46.41	0.4388	10.910
125-50	Low	292	57.7	184.01	44.95	0.2284	5.598

Common Pb ranged from 10 to 20%; correction used:
1: 13.52: 14.639: 33.261

1 Observed IR values

Barnston River the metasediments are less metamorphosed than to the west, where coarse-grained "paragneisses" are present, generally intruded by late granite sheets and showing local evidence of partial melting. Units termed "gneissic complexes", which occur near contacts with granitoids and separately, consist of the local paragneisses with intrusive granite sheets and veins. The metasediments (=paragneisses) are clearly derived from turbidites and are not part of an early basement. Chemically, they have moderately high SiO_2 (60-70%) and relatively high MgO. Coarser and finer-grained components of the turbidite sequence have an overlapping range of compositions and cannot be distinguished on the basis of major element chemistry. Furthermore, whether in the eastern or western belts, or whether intruded thickly by granite sheets or not, the paragneisses show no detectable local variations in composition. The only known cause for the range of compositions lies in hydrothermal alteration of biotite. A subset of the aluminous "Paragneiss 2" has $\text{K}_2\text{O} < \text{Na}_2\text{O}$, reflecting chloritization of biotite. No other chemical changes can yet be proved to accompany this loss of K. So far, it appears that the chemistry of the paragneisses is governed more by physical than by chemical sedimentary processes during transportation and deposition, as coarser and finer fractions appear uniform.

The intrusive granite sheets and isolated plutons are mostly hololeucocratic two-mica granites. Their Rb-Sr age is estimated as 2540 ± 10 Ma, $R1 = 0.702$, probably close to the true emplacement age and therefore of the age of the regional metamorphism.

Zircons from one block of paragneiss show no evidence of the presence of an ancient basement. The average age of the source of the zircons appears to be 2540 Ma, close to the age of the Yellowknife Greenstone Belt and the Western Granodiorite at Yellowknife.

ACKNOWLEDGEMENTS

This project was carried out under the terms of a contract with the Department of Indian Affairs and Northern Development; the laboratory work was financed by NSERC (grant A-7489) and by NERC (U.K.), all of which assistance is gratefully acknowledged. Advice on zircon work was freely given by Dr. H. Baadsgaard, to whom our thanks are due.

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MINERALOGY OF THE CO-PRODUCT METALS IN THE
T-ZONE BERYLLIUM DEPOSITS, THOR LAKE, N.W.T.

by

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The Thor Lake T-zone rare-metals deposits, located 95 km southeast of Yellowknife, contain economically significant concentrations of Be, Y, REE, Nb and Ga related to albitization and greisenization of peralkaline granite and syenite. The T-zone contains four lithologic units including: the Wall zone, the Lower Intermediate zone, the Upper Intermediate zone and the Quartz zone. Beryllium is the principal ore metal in the T-zone deposits. The purpose of this research project is to examine in detail the mineralogy of the co-product metals. Specifically, this study is concerned with the identification and nature of the various Y-, REE-, Nb- and Ga-bearing minerals present and their distribution within the beryllium deposits. Of particular interest are the sizes, textural relationships and gangue mineral associations of these potentially economic minerals. Standard petrographic techniques and electron microprobe analysis in conjunction with assay data are being used in this work.

By December, 1984, a preliminary study of samples from the Lower Intermediate, Upper Intermediate and Quartz zones had resulted in the tentative identification of seven separate Y-, and REE-bearing minerals. The results of this initial work were presented at the 12th annual Geoscience Forum in Yellowknife as a talk entitled "Y and rare earths in the Thor Lake rare-metals deposits". Subsequent work conducted on samples from all four lithologic units has provided information on the mineralogy and distribution of Nb and Ga in addition to Y and REE. Associated Be-bearing minerals have also been studied briefly.

To date three Y-bearing minerals have been identified. Gadolinite ($Y_2FeBe_2(SiO_4)_2O_2$) and xenotime (YPO_4) appear to be the most abundant species. Textural relationships suggest that where the two occur together gadolinite is replacing xenotime. Both are found in the Lower Intermediate and Upper Intermediate zones, although xenotime also occurs as inclusions in quartz in the Quartz zone. The third Y-bearing mineral is an unidentified Th,Fe,Y-silicate that is radioactive and may possibly be a new mineral species. Volatilization effects under the electron beam during microprobe analysis suggest that this mineral may be hydrous. All three Y-bearing minerals are found predominantly in the Lower Intermediate - Upper Intermediate zones.

The most abundant REE-bearing minerals are three LREE-enriched fluorocarbonates tentatively identified as bastnaesite ($Ce(CO_3)F$) and two of the three Ca-bearing phases: synchisite, parisite and roentgenite. These phases occur as normal intergrowths with other minerals and as syntaxial intergrowths with two, and possibly all three, present together in one polycrystal. These fluorocarbonates are found throughout the deposit in all four lithologic units. Monazite ($(Ce,La,Y,Th)PO_4$), another LREE-enriched

phase, is found in all rock types excluding the Quartz zone. High La, Ce and Nd assays are attributed to the presence of one or more of these LREE-bearing minerals. High Gd assays are attributed to gadolinite and xenotime.

Concentrations of Nb minerals occur predominantly in the Wall zone and to a lesser extent in the Lower Intermediate zone. The principal Nb-bearing mineral in the T-zone deposits is a Nb oxide, probably columbite ($(Fe,Mn)(Nb_2O_6)$) or ashanite ($(Nb,Fe,Mn)_4O_8$). In the Wall zone, the Nb oxide occurs as inclusions in quartz and alkali feldspar and very rarely forms small veinlets. In the Lower Intermediate zone, discrete grains and larger aggregates of the Nb oxide are found as inclusions in quartz and biotite (?). A second Nb-bearing mineral, associated with the Nb oxide in the Lower Intermediate zone, is an unidentified Nb,Ti,U-silicate. Volatilization effects under the electron beam suggest that this mineral may also be hydrous. Tantalum in these Nb-bearing phases is present in trace amounts only.

Ga-bearing minerals are found exclusively in the albitized rocks of the Wall zone. Assay results indicate an average content of 500 ppm Ga in the Wall-zone rocks. The Ga occurs in albite laths where it is probably substituting for Al^{3+} . Preliminary analytical work has shown considerable variation in the distribution of Ga from grain to grain within a single rock and from rock to rock within the same lithologic unit. Further work is required to clarify the distribution of Ga within the albite laths.

Associated Be-bearing minerals are phenacite (Be_2SiO_4), bertrandite ($Be_4Si_2O_7(OH)_2$), gadinolite (as mentioned above) and a member of the helvite group ($(Mn,Fe,Zn)_4Be_3Si_3O_{12}S$).

The results of research completed to date will be presented at the 1985 GAC/MAC meeting in Fredericton as a poster session entitled "Economic mineralogy of the T-zone rare-metals deposits, Thor Lake, N.W.T."

Continued research towards an M.Sc. thesis on the mineralogy of the co-product metals in the T-zone deposits will include:

1. examination of remaining samples from the four mineralized lithologies and unmineralized host rock to complete the mineralogical study; determining the presence, nature and distribution of co-product metals.
2. completion of quantitative analyses on:
 - a) the Y-, REE- and Nb-bearing minerals for positive mineral identification
 - b) the Ga-bearing albite to determine the distribution and concentration of Ga within single mineral grains.

The tentative date of completion of the M.Sc. thesis is December, 1985.



**GEOCHEMISTRY, PHASE RELATIONS, AND GEOTHERMOMETRY
AS INDICATIONS OF PRIMARY ROCK HISTORY
AND CONDITIONS OF METAMORPHISM
AT RUTLEDGE LAKE, N.W.T.**

by
M. Fyson
Ottawa University

ABSTRACT

Rutledge Lake, N.W.T., lies on a north-trending shear belt in the west-central Churchill Structural Province. The belt appears on magnetic anomaly maps as a steep inflection which continues north to the McDonald Fault where it curves in a dextral sense NE along the fault zone. The magnetic signature and sense of shear are similar to those found along the western edge of the Thelon Tectonic Zone (Thompson and others, 1985) northeast of the fault.

Six units of granulite-facies rocks at Rutledge Lake can be discriminated into granites, mafic to ultramafic meta-igneous rocks and metasediments. These are classified using whole-rock geochemistry as calc-alkaline granites, tholeiites, basaltic komatiites and sediments that possibly include Slave provenance turbidites.

Equilibrium assemblages in pelitic and basic rocks (eg. cordierite-garnet-sillimanite; Ca-pyroxene-hypersthene-plagioclase) and reactions derived from observed phase relations outline high-T (960 ± 60 degrees Celsius), low-P (3.5 to 5 Kb) metamorphic conditions probably caused by intrusion of the granites.

INTRODUCTION

In this paper I will attempt to constrain the conditions of metamorphism and determine the protoliths of some granulite-facies rocks from Rutledge Lake, Northwest Territories. Mineralogical phase relations (postulated from petrographic observations) and mineral and whole-rock chemical analyses will be used to unravel the metamorphic history; trace and major element geochemistry and relict field morphologies of units will give an indication of the pre-metamorphic geology of the area.

I would like to express my gratitude to Nick Culshaw and Ralph Kretz for their help and encouragement throughout this project. Funding for laboratory work was provided by the University of Ottawa and by H. Bostock of the Geological Survey of Canada.

GENERAL SETTING

The study area, Rutledge Lake, lies along approximately 30 km of a north-trending shear zone in the west-central Churchill Structural Province about 50 km south of the East Arm of Great Slave Lake, N.W.T. (Fig. 1). The shear zone coincides with a magnetic inflection which runs north into the McDonald Fault (GSC Aeromag Map no. 1566A). This inflection appears to be sinistrally offset by the McDonald Fault, and appears to continue north of the fault as a similarly offset inflection coinciding with the western edge of the Thelon Tectonic Zone, an inter-plate (Slave-Churchill) boundary area or intra-plate shear area (Thompson and others, 1985). The Rutledge Lake shear zone may therefore represent either an inter- or an intra-plate shear zone.

Mapping by Culshaw (1984) at a scale of 1:30,000 has distinguished several ortho- and paragneissic units. For the purpose of this study these are divided into the six lithological units described below and shown in Figure 2.

Units 1a and 1b: Rutledge Lake Complex

A central granite and a central metasedimentary unit make up what is termed the Rutledge Lake Complex. This is an interlayered ortho- and para-

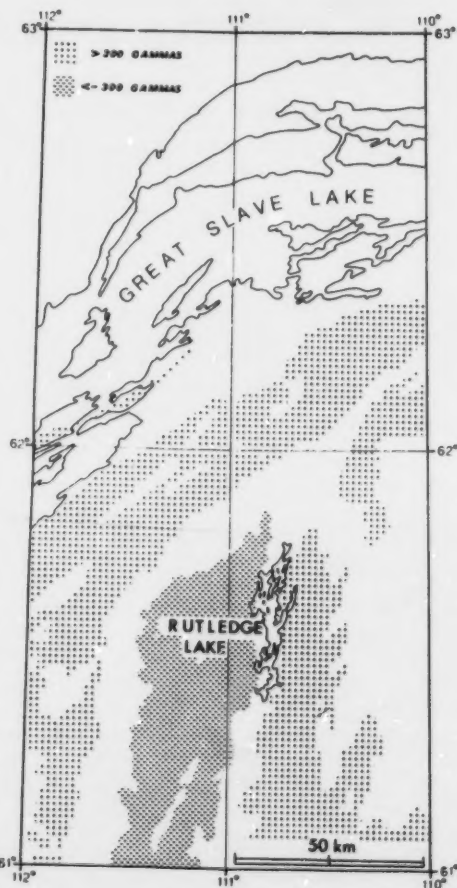


FIGURE 1: Index map of study area centered on Rutledge Lake. The magnetic anomaly pattern is from GSC map 1566A (from Culshaw, 1984).

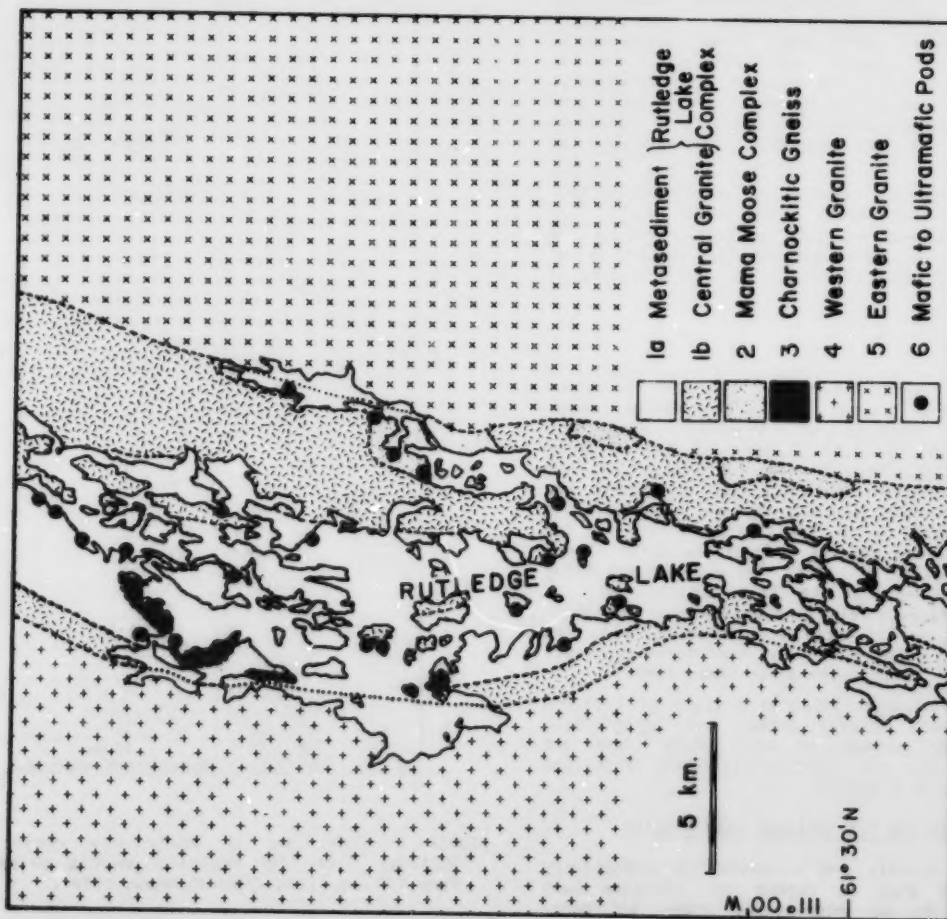


FIGURE 2: Geology of Rutledge Lake.



FIGURE 3: Photo showing variation of garnet porphyroblasts (dark specks) in quartzite (lighter layers) and pelitic (darker layers) of the Rutledge Lake Complex. Repeated coarsening upward of garnets is interpreted as inverse grading mimicking primary fining upward sequence in turbidites.

gneiss with layering on the decimeter to hectometer scale. Paragneiss predominates towards the center of the complex and orthogneiss towards the outside edges. The metasediments are interlayered bluish- to white-weathering quartzites and grey-brown-weathering pelites, the latter commonly carrying medium to coarse sillimanite and garnet porphyroblasts. Gossanous zones characterized by the presence of pyrite and graphite are common within the metasediments. Blue quartz is characteristic of the whole Rutledge Lake Complex. It is often stretched into ribbons highlighting the north-trending, steeply-dipping and shallowly-plunging LS fabric centered along the Rutledge Lake Complex. The granitic component cuts the metasedimentary component and includes within it xenoliths and xenocrysts of metasediment at all scales. Figure 3 is a photograph of a fairly undeformed outcrop of the Rutledge Lake Complex metasediments, showing the relation between pelitic and quartzitic members. This is the only locale where repetitive and continuous decimeter-scale layering can be clearly identified since other outcrops are either migmatized with granitic rock or are massive quartzite or pelite. Figure 4 shows the outcrop-scale characteristics of the central granite. The horizontal trace of the LS fabric and some small shears offsetting a mafic band are visible.

Unit 2: Mama Moose Complex

Along the eastern margin of the central granite are tracts of metasedimentary rock interlayered with mafic granulite rock known collectively as the Mama Moose Complex. The metasediments are bluish- to pinkish-weathering rocks, generally uniform in

contrast to the Rutledge Lake Complex and not cut by granite. Abundant magnetite and sillimanite porphyroblasts are characteristic of the unit. Garnet and cordierite are not abundant. This unit shows up as a distinct positive anomaly on the geomagnetic maps.

Unit 3: Charnockitic Gneiss

Within the Rutledge Lake Complex, towards the north of the lake, is a body of charnockitic gneiss. This is a distinctive, fine-grained rock weathering rusty-buff to blue-grey and quite dark green on the fresh surface. Hypersthene and (rarely) garnet are visible in hand specimens.

Unit 4: Western Granite

The granitic component of the Rutledge Lake Complex grades into this unit to the west. It is a medium-grained, pale pink granite, and although sedimentary xenoliths are not abundant, it contains aluminosilicate phases such as spinel, garnet and sillimanite. It is coarser grained, less recrystallized and less deformed than the central (Rutledge Lake Complex) granite.

Unit 5: Eastern Granite

This unit is composed mainly of a fine-grained, uniform, pink granite. In some areas a grey, strongly-deformed and recrystallized formerly megacrystic granodiorite has been included. Two features of the Eastern Granite are inclusions of black, decimeter- to meter-scale amphibolite pods, and a N-S striking shear system which mylonitizes the rock to the east.



FIGURE 4: Outcrop characteristics of Central Granite (Rutledge Lake Complex). Note horizontal trace of LS fabric highlighted by ribboned quartz. Minor shears offset a mafic layer. Pencil for scale.

Unit 6: Mafic to Ultramafic Pods

These range from about 1 to 50 meters across, and are found within the Rutledge Lake and Mama Moose complexes and in the Eastern Granite. Most are amorphous blobs, but one is sill-like. The differing field relations of these rocks in the two complexes are discussed by Culshaw (1984). Figure 5a shows a pod of ultramafic rock invaded by granite. The contrasting appearances of rocks classified in the field as 'mafic granulites' and 'ultramafics' are seen in Fig. 5b. Clinopyroxene is visible in the 'ultramafic' member.

PETROGRAPHY

About fifty thin sections of metamorphic rocks from the Rutledge Lake area were examined. Sericitic and chloritic alteration is common and sometimes pervasive, but is minimal in many sections. In a few sections alteration to epidote is observed.

Three main metamorphic rock types are present in the Rutledge Lake area; derived from granitic, ultramafic to mafic igneous, and sedimentary parent rocks respectively. Petrographic descriptions of the individual units have been summarised by Culshaw (1984).

Granitic Rocks

The mineralogical characteristics of the granitic rocks are quite close to that of unmetamorphosed granites: quartz-feldspars-opaques $\pm$ biotite. A typical thin-section looks like Fig. 6a, with quartz either as rounded or stretched-out polygranular grains and feldspars often showing mortar fabrics. The western and central granitic rocks have xenomorphic, corroded-looking inclusions (probably xenocrysts) of garnet, spinel, sillimanite, and cordierite -- the latter altered to chlorite and sericite. These are absent in the eastern granitic gneisses. The charnockitic gneiss is a granuloblastic hypersthene-bearing variety with biotite, magnetite, orthoclase, plagioclase and mesoperthite (Fig. 6b).

Mafic to Ultramafic Rocks

The mafic to ultramafic rocks are interesting because of their mineralogy and mineralogical associations. The 'mafic granulite' samples (mafic) show granuloblastic (equilibrium) textures with the assemblage orthopyroxene-clinopyroxene-plagioclase-opaques $\pm$ biotite. Figure 6c is a photomicrograph of a mafic granulite sample.

Not many of the 'ultramafics' were examined in thin section. Those that were show two-pyroxene (OPX-CPX) or olivine-orthopyroxene assemblages $\pm$ biotite $\pm$ spinel (Fig. 6d). These rocks in some localities at Rutledge Lake are related to copper-nickel sulphides.

Although equilibrium textures are not found in the ultramafic rocks, both these and the mafic rocks have two-pyroxene assemblages. This makes them good candidates for geothermometry studies.

Metasediments

The metasedimentary rocks have mineral assemblages that can be used to determine conditions of metamorphism. A description more detailed than for the other rock types is therefore required. Phase

relations observed in these rocks are used in a later chapter to put constraints on the conditions of metamorphism. A thin section of typical cordierite- and sillimanite-bearing metasediment is shown in Fig. 6e.

Sediments of the Rutledge Lake Complex vary from quartzites to pelites showing varying parts of the assemblage: quartz-garnet-cordierite-sillimanite-feldspar (K-feldspar, plagioclase or perthitic intergrowth)-hercynite spinel-opaques (magnetite or ilmenite with hematite exsolution)-biotite. Epidote and zircon are also present, as are the alteration products sericite and chlorite. All minerals are xenomorphic to subidiomorphic. Quartz occurs as strained flasers and ribbons up to 1 cm long (sometimes longer), with seriate amoeboid boundaries, and also as fine (0.2 mm) recrystallised grains. It is also found as inclusions in garnet porphyroblasts. All phases present show sutured boundaries, but triple junctions are present. In some areas where orthogneiss and paragneiss are intermixed, feldspars are found only in the granitic component of the rock.

Sediments of the Mama Moose Complex are mapped in four separate localities to the east of Rutledge Lake (Fig. 2). Samples were studied from the two northernmost localities. There is some variation in mineralogy between these two areas. The northernmost Mama Moose paragneiss shows an assemblage of quartz-sillimanite-cordierite-opaques-feldspar (plagioclase, K-feldspar or perthitic intergrowth), with sericite and chlorite as possible retrograde minerals. No garnet and little biotite is present. Further south another body of paragneiss of the same unit shows the same assemblage as above, but with more biotite. At the south end of this body, garnet is added to the assemblage. The predominant mineral species in these paragneisses often show sutured contacts which do not indicate that equilibrium has been reached. Quartz occurs as flasers less or equal to 3 mm long with some mortar fabric. Grains of all mineral species are often partly or totally surrounded by sericitic alteration. Cordierite alteration to sericite commonly consumes the cordierite completely. Sillimanite is common throughout in bundles up to 5 mm across, and in single needles up to 3 mm long.

CHEMISTRY

Bulk rock chemistry was determined by X-ray fluorescence analyses of 23 samples drawn from all of the rock types in the area. Sample locations are shown in Figure 7. Samples were chosen for their distinctive and characteristic field appearances and mineralogies. This work was supplemented by trace element analysis of Y, Cr, Zn, and Cu on a Direct-Current Plasma (DCP) spectrograph. Because parental rock classifications are quite confidently distinguished in outcrop, I will refer in the following discussion to the primary rock names of the units analysed. Table 1 explains the sample numbering scheme used on all of the diagrams, and correlates the plotted numbers to the field station-sample-thin section identifications.

The analysed samples are first plotted on an AFM projection taken from Reinhardt, 1967 (Fig. 8). Though this diagram is designed to represent phase assemblages found primarily in pelitic sedimentary rocks, it illustrates distinctions between major element chemistries of the sediments, granitic, and mafic to ultramafic rocks. The highly alkalic nature of the granites, the low-Mg characteristic of the



FIGURE 5a: Typical ultramafic pod injected by granite.



FIGURE 5b: Ultramafic pod in mafic granulite. Both olivopyroxene and orthopyroxene are present in the ultramafic pod.



FIGURE 6a: Thin section of typical granite in central Rutledge Lake shear-zone area. Ribboned material is quartz. Plagioclase, K-feldspar and biotite are also present. Crossed polars. Photo is 5 mm along edge.

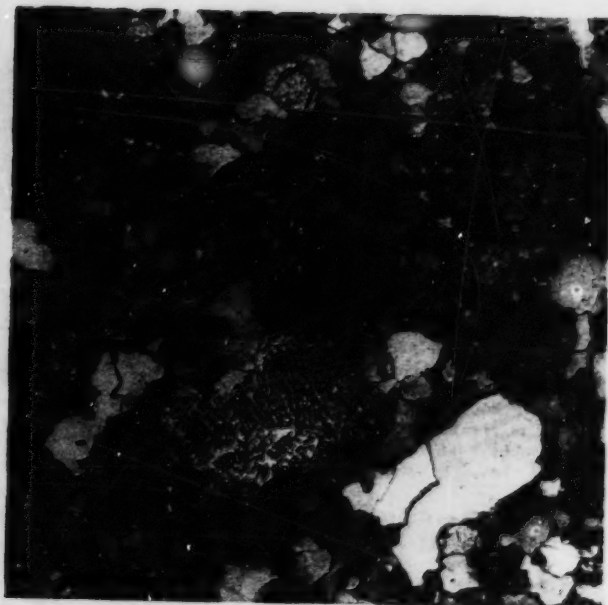


FIGURE 6b: Thin section of charnockitic gneiss (sample #189) viewed under crossed polars. Plagioclase, K-feldspar, mesoperthite, orthopyroxene and opaques are present. Photo 7 mm along edge.



FIGURE 6c: Thin section of tholeiitic mafic granulite (sample #180) viewed under crossed polars. Note granoblastic mosaic of clinopyroxene, orthopyroxene and plagioclase. Photo 4 mm along edge.



FIGURE 6d: Thin section of ultramafic komatiite (sample #197a) viewed under crossed polars. The higher-relief mineral is olivine and the lower-relief mineral is orthopyroxene. Photo 2 mm along edge.

TABLE 1: SAMPLE IDENTIFICATION SCHEME

The following legend applies to all Figures and Tables (including the sample location map and the table below):

▲ Granites (Granitic gneisses)	WG	West Granite
□ Metasediments (Pelites)	CG	Central Granite
○ Ultramafic-Mafic Igneous Rock	EG	East Granite
	CH	Charnockite
	RL	Rutledge Lake Pelite
	MM	Mama Moose Pelite

Sample Number	Rock Unit (Granitic Rx)	Sample Number	Rock Unit (Sediments)	Sample Number	Enclosing Unit (Mafic-Ultramafics)			
* 1	257b	EG	* 1	53a	RL	** 1	180	MM
2	114	CG	2	93b	MM	2	96c	CG-RL
3	169	CH	3	171a	RL	3	238b	CG-RL
4	61a	CG	4	62b	MM	4	279b	CG-RL
5	196	WG	5	142 1a	MM	5	263	MM
						6	76b	CG-MM
						7	174	CG-RL
						8	127c	CH-RL
						9	270b	MM
						10	266b	MM
						11	105a	CG-RL
						12	197	CG-RL

* In order of increasing SiO_2 content.
 ** In order of increasing MgO content.

For the Henderson (1972) compilation plots shown in Figs. 8 and 9, the legend is as follows (all are averages):

- A 3 Burwash Formation greywackes (Slave Province)
- B 20 Archean greywackes
- C 61 All-age greywackes
- D 3 Burwash Formation mudstones (Slave Province)
- E 20 Archean slates
- F 36 Phanerozoic slates
- G Average comp. of the Canadian Shield
- H Average granodiorite



FIGURE 6a: Thin section showing typical pelite viewed under crossed polars. Cordierite is twinned. Sillimanite is the long crystal curving around the cordierite. Quartz, feldspar, garnet, biotite and opaques are also present. Photo 2 mm along edge.

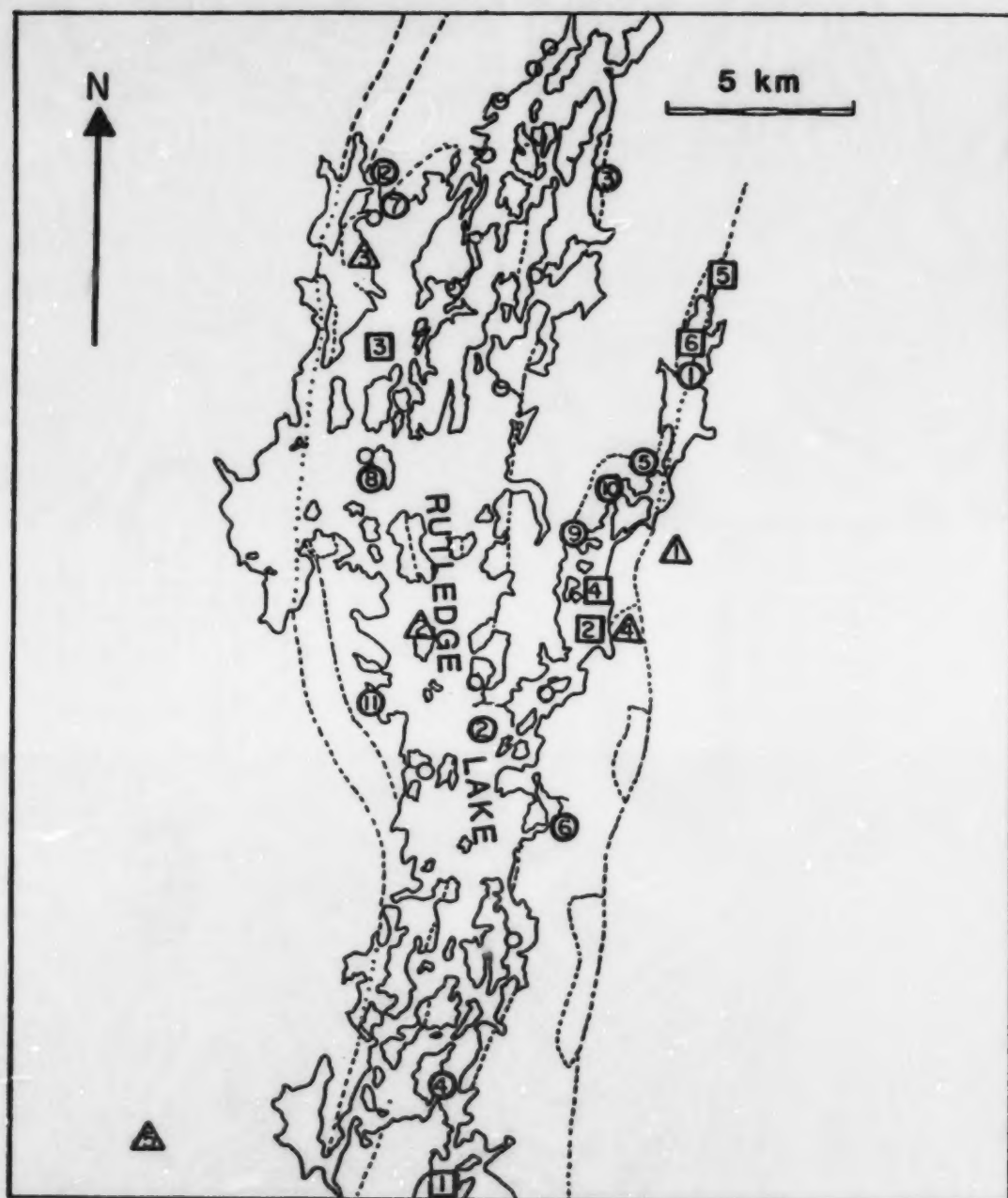


FIGURE 7: Geochemical Sample Locality Plan. $\triangle$ - granites, $\square$ - metapelites, $\circ$ - mafic to ultramafic meta-igneous rocks.

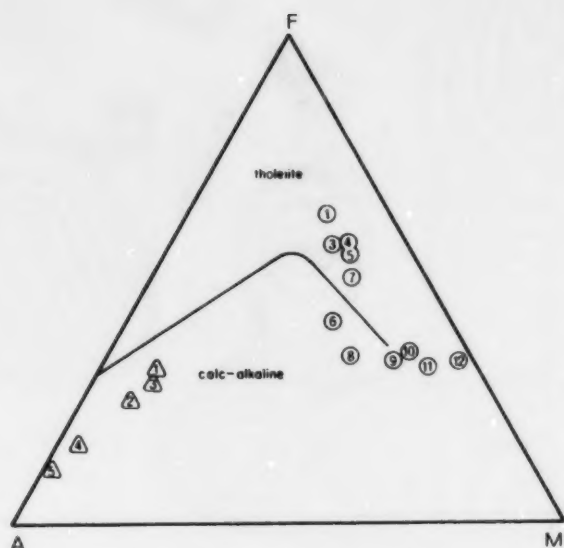


FIGURE 8: AFM Diagram (from Reinhardt, 1968).
 $A = Al_2O_3 + K_2O + Na_2O + CaO$; $F = FeO + Fe_2O_3 + TiO_2$; $M = MgO$. Solid squares (A to H) are from Henderson's 1972 compilation for greywackes and mudstones (see Table 1).

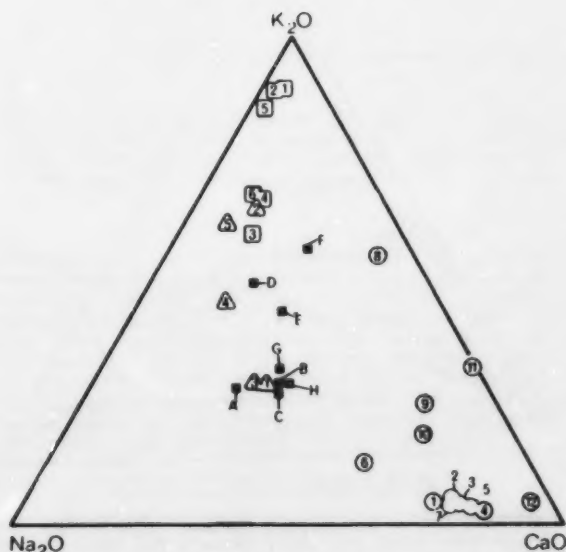


FIGURE 9: CNK alkalis diagram. Solid squares (A to H) are from Henderson's 1972 compilation for greywackes and mudstones (see Table 1).

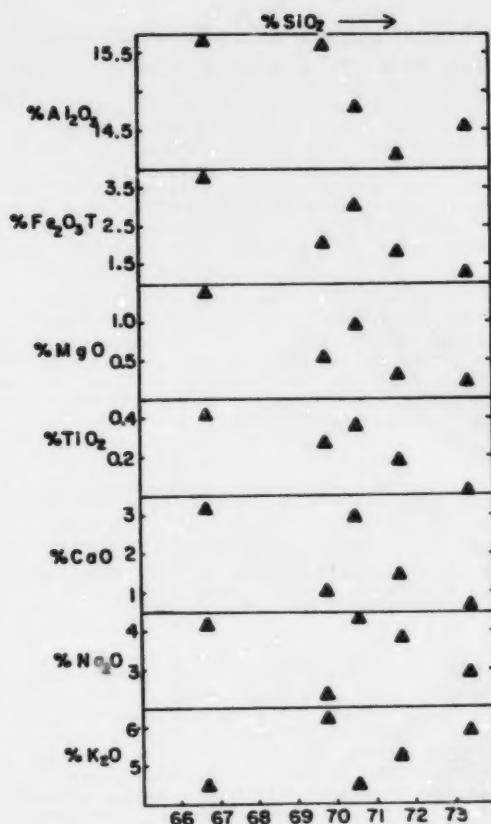


FIGURE 10: Harker variation diagrams for the granitic rocks. Sample numbers are 1 to 5 going from left to right.

pelites, and the higher Mg-Fe-Ti content of the mafic to ultramafic rocks with respect to the oxides represented is apparent.

The alkalis are plotted on Figure 9 - a CNK plot. A fairly constant Na:Ca ratio is noted in the granites and pelites. The pelites show higher K relative to Na and Ca than the granites. The mafics to ultramafics are of course higher in Ca relative to Na and K. Harker diagrams shown in Figure 10 show abundances of these oxides in relation to SiO_2 for the granitic rocks.

Having established the major-chemistry differences between the rock types of the Rutledge Lake area, I will now try and make more specific genetic interpretations for the three main rock types.

Granites

Both normative analyses (2) and modal counts (5) of the granite specimens chosen are plotted in Figure 11 on Streckheisen's Quartz - K-Feldspar - Plagioclase granite classification plot, which is designed for modal analyses. The normative analyses are low in quartz compared to the modal counts. The formation of (low-silica) sericite, which is common in the

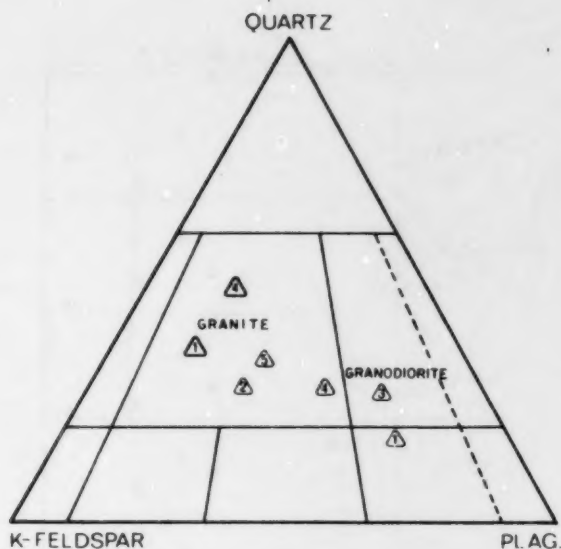


FIGURE 11: IUGS granitic rock classification diagram. Heavy triangles = from modal counts; light triangles = from normative analyses.

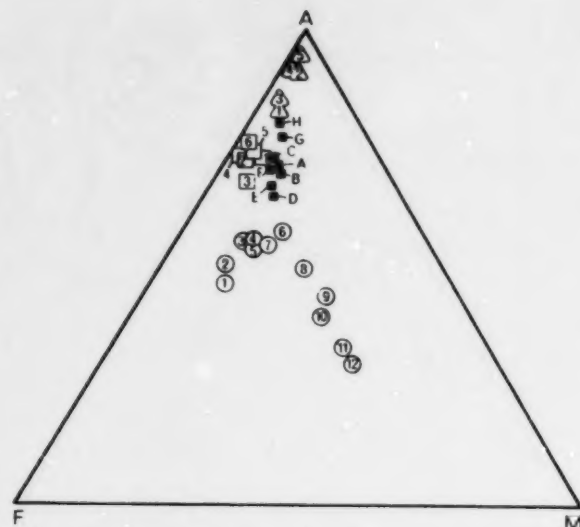


FIGURE 12: AFM plot with Irvine-Baragar dividing line between tholeiitic and calc-alkaline suites. $A = \text{Na}_2\text{O} + \text{K}_2\text{O}$; $F = \text{FeO} + \text{Fe}_2\text{O}_3$; $M = \text{MgO}$

granites but was not point-counted, could be responsible for boosting modal quartz content relative to feldspars. The sericite may be a cause of both analytical and statistical error in determining the modal plots of Fig. 11. The granites are therefore shown to lie in the granite-granodiorite fields in this classification scheme. Plotted in Figure 12 on an alkali-iron-magnesium (AFM) plot, the granites are shown to follow a calc-alkaline trend. Note the nice linear trend of the plots in this diagram. Harker plots of oxide proportions relative to SiO_2 for the granites (Fig. 10) show similar linear trends of decreasing Fe, Mg and Ti oxides with increasing SiO_2 going from the eastern through central and western granites. Plots of Al, Ca, Na, and K oxides show an inverse relationship of K and Na+Ca.

For the three variation diagrams which appear to describe clear trends (Fig. 10), it is interesting to note that the charnockitic gneiss (sample 3) plots off the apparent trends (more Fe, Ti, and Mg-rich with respect to SiO_2). The significance of this is unclear. On the triangular diagrams used in this study to examine the granites, granite samples 3 and 2 often seem to be reversed in position. Combined with the observations recorded from the Harker diagrams, this might suggest that the charnockitic gneiss sample is SiO_2 -enriched relative to its mafic mineral content, in the order of about 2% SiO_2 by weight. By adjusting the silica content of sample 3 as described above, the granites show a trend of decreasing Ca and Na and increasing K with increased silica, with the exception of sample 2, a particularly K-rich and Ca-Na-poor rock. One can also define a general decrease in Al with increase of silica. Samples 2 and 5, both high in Al with respect to this trend, are from the two granite bodies which contain aluminosilicate phases.

The apparent correlations of the granitic suites discussed here may be expected as natural trends for acid igneous rocks of this type or may demonstrate

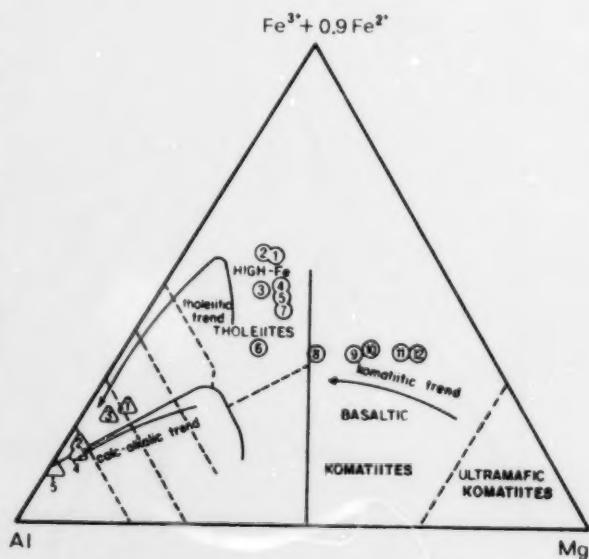


FIGURE 13: Jensen cation plot (Jensen, 1976).

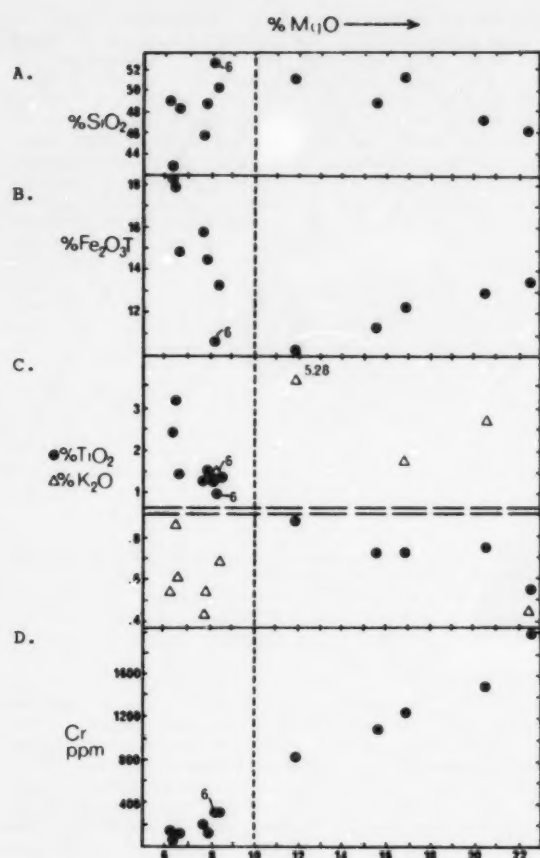


FIGURE 14: Variation diagrams for mafic to ultramafics. The tholeiitic rocks plot to the left of the short-dash line; the basaltic komatiites to the right. Samples are numbered from 1 to 12 from left to right.

a progressive generation of each of the plutons represented, for example, by magmatic pulses from an underlying source such as a batholith. Either explanation points to the conclusion that in spite of the amount of sedimentary material included in the granites, they are a primary magmatic feature related to the sediments by assimilation of the latter and not by mass melting of the sediments. The two rock types are also, as previously stated, chemically distinctive from one another.

Mafic to Ultramafic Igneous Rocks

Geochemical analyses of the mafic to ultramafic rocks on the AFM diagram (Fig. 12) shows that there are two groupings of rock chemistries with one sample transitional between the two. Five samples (8 to 12) plot as more Mg-rich, perhaps less differentiated intrusions or flows. A further six samples appear to be (more differentiated?) high-Fe tholeiitic rocks. All are fairly alkaline-poor with respect to Fe and Mg, and on the CNK diagram (Fig. 9) have high Ca with respect to K and Na. Note the single, exceptionally

potassium-rich mafic to ultramafic sample (#8).

When the data is plotted on the Jensen Cation Diagram (Fig. 13) one can distinguish two types of mafic to ultramafic rocks. The higher-Mg rock types show a well-defined komatiitic trend; they can be classified as basaltic komatiites. The higher-Fe-bearing rocks lie along a tholeiitic trend and are high-Fe tholeiitic basalts. The term basalt may be taken to have extrusive genetic implications, however since the mafic to ultramafic rocks at Rutledge Lake may be intrusive or extrusive, no such genetic interpretation is implied.

In the variation diagrams of Figure 14, Cr and the oxides of Si, Fe, Ti and K are plotted against MgO. The diagrams illustrate some of the important differences between the two mafic to ultramafic rock types, as well as showing some important chemical trends. Silica variation is shown to demonstrate that the field classification of 'mafic-ultramafic' (which usually corresponds with the chemical classification 'basic-ultrabasic' based on silica content) cannot be used to distinguish tholeiites and basaltic komatiites at Rutledge Lake. Both rock types span a similar range of silica content. It is interesting to note a statistical increase in silica content in the tholeiites and decrease in silica content in the basaltic komatiites with increase in MgO content. The reverse of these trends are shown clearly for Fe vs. MgO (Fig. 14b).

Figure 14c shows that with one exception for each rock-type, the basaltic komatiites are high and the tholeiites are low in potassium. The tholeiite exception is the same sample (#6) which shows up on the AFM plot as being transitional between the two rock types. In all of the diagrams of Figure 14 this sample appears to closely correspond to the trends shown by the basaltic komatiites, implying that it is related to the basaltic komatiite suite rather than to the tholeiites. The komatiitic exception, perhaps coincidentally, is from an ultramafic pod that yielded high Cu-Ni assays.

There is a statistical increase in chromium content with increasing MgO in both rock types (Fig. 14d), but the relationship is much clearer and more distinct in the basaltic komatiites. Chromium content is an order of magnitude greater in the basaltic komatiites than in the tholeiites.

Linearly increasing or decreasing relationships of element proportions as described above may result from unrelated magmas with close starting compositions undergoing similar fractionation trends (therefore implying that the rocks may not be genetically related) or may be attributed to cogenetic processes (for example progressive magmatic differentiation of a single magmatic source). It is more likely (less coincidental) to assume the latter in very linear trends such as those shown by the komatiitic rocks, however more scattered plots like those shown by the tholeiitic rocks are less convincing of a cogenetic origin. However, the inverse relationship of iron to magnesium seen here in the tholeiites is quite common worldwide in granitic rock considered by most researchers to have formed by cogenetic magmatic differentiation.

Low Ti and high Cr in the basaltic komatiites are quite compatible with published compilations (Arndt and Nisbet, 1982), and the trend of Ti decreasing linearly with increase in MgO is seen in almost all compilations of unrelated flows (hence this is probably an inherent characteristic of komatiitic

rocks). However, the nice linear relationship of chromium to magnesium is not implicit in the nature of komatiitic rocks (which show a fair scatter of points in compilations of different flows, see Arndt and Nisbet, 1982).

Both Cr and Mg are highly compatible in regard to mantle composition, and accordingly both increase similarly in the liquid phase in proportion to increasing degrees of partial melting (R. Baragar, pers. comm.). A linear relationship shown by the Rutledge Lake komatiitic suite is therefore a good indication of a cogenetic relationship between the komatiitic rocks.

High potassium concentration in komatiitic rocks is unusual (Arndt and Nisbet, 1982). This may be a metamorphic effect since potassium is a very mobile component during metamorphism, however, if this were true one might expect that the same effect would be seen in the tholeiitic rocks. An examination of Appendix 1 reveals an absence of any potassium-bearing phases in the tholeiitic rocks as compared with a relative abundance of these phases (micas, K-feldspar) in all other rocks of the Rutledge Lake area. This appears to account for the present low-K chemistry recorded in the tholeiites.

All the mafic to ultramafic rocks analysed were plotted on the Ti-Zr-Y trace-element discrimination diagram of Pearce and Cann, 1973 (Fig. 15). The results are too scattered to draw any useful observations or conclusions. However, high titanium is characteristic of an inter-continental origin for the tholeiites (R. Baragar, pers. comm.).

Pelitic Sediments

On the AFM projection (Fig. 8) the chemistry of the pelites is compared to the chemical averages of greywackes and mudstones from the Slave and other geological provinces compiled by Henderson (1972). A slightly lower Mg-content with respect to the other major elements is recorded for the Rutledge Lake pelites. In the CNK plot (Fig. 9) one can see that the potassium content with respect to the other alkalis is somewhat higher in the Rutledge Lake area pelites than in the Henderson compilation rocks. Subtracting 10-15% K from the Rutledge Lake area pelites gives the same $\text{CaO-Na}_2\text{O-K}_2\text{O}$ ratio as the Burwash Formation samples plotted by Henderson. Considering that the rocks of the Henderson compilation have been metamorphosed only to low grade while the Rutledge Lake area pelites are high-grade granulite-facies rocks, and recognising that potassium is fairly mobile as a metamorphic component, the potassium variability may be due to introduction either by metamorphism or by intrusion and contamination by the K-rich granites. Magnesium variability is most likely an effect of source-rock chemistry. The fairly constant Na:Ca ratio previously noted in the granites and pelites from Rutledge Lake is also noted in the Henderson compilation rocks. Thus, aside from differences in potassium and magnesium content, possibly accounted for in the explanations above, I feel that the pelitic members of the Rutledge Lake and Mama Moose sedimentary complexes have major-element chemistries comparable to Slave Province mudstones.

Field data gathered from a relatively undeformed outcrop from within the Rutledge Lake complex sediments provides further evidence that these rocks include water-lain mass-flow deposits. At this locale, quartz-rich layers with very fine garnet

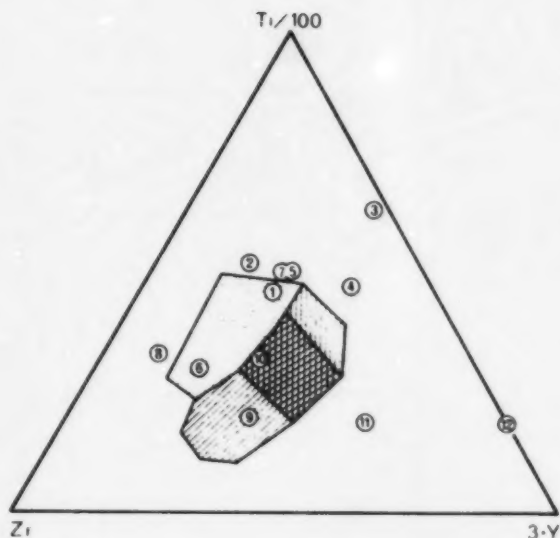


FIGURE 15: Trace-element discrimination plot for basic volcanics (from Pearce & Cann, 1973). Stippled area = within-plate basalts; Diagonal stripes = calc alkaline basalts; Vertical stripes = low potassium tholeiites.

porphyroblasts grade one direction into pelitic layers with coarser and more abundant garnets (Fig. 3). Continuing in the same direction across the layers, the pelitic members become rich in sillimanite before terminating in abrupt contacts against the next quartz-rich bed. This inverse grading of porphyroblasts, which I have seen elsewhere in the Slave Province, mimics the primary layering in metamorphosed turbidite greywackes.

Outcrops of pure pelite and pure quartzite around the Rutledge Lake Complex may indicate varying depositional environments or different facies (distal and proximal) of turbidites.

CONDITIONS OF METAMORPHISM

The two main types of data used to determine the conditions of metamorphism in the rocks at Rutledge Lake are the petrological phase relations observed in the metasedimentary rocks and the application of the two-pyroxene geothermometer in the mafic to ultramafic rocks.

The coexisting mineral phases characteristic of the Rutledge Lake Complex metapelites, which are useful in determining the conditions of metamorphism are: sillimanite-almandine garnet $\pm$ cordierite $\pm$ biotite $\pm$ hercynite spinel; in the Mama Moose Complex they are: sillimanite-cordierite $\pm$ almandine garnet $\pm$ biotite. Quartz, feldspars, and magnetite or ilmenite are present throughout.

In thin-section, all mineral phases of the equilibrium assemblages (following diagrams) are in mutual contact, though equilibrium textures (for example, crystalloblastic mineral intergrowth) are not characteristic.

The studies of Reinhardt (1967) show that the bulk composition of a rock should lie within the

triangle connecting the equilibrium assemblage minerals found within it. The bulk compositions of the metapelites from the Rutledge Lake area are plotted on an AFM Diagram (Fig. 16). Tie-lines connect the phases shown on the diagram which coexist as equilibrium assemblages in the rocks. Crossed tie-lines require that two variations of this diagram be constructed. Fig. 17a has almandine garnet and cordierite connected by a tie-line and Fig. 17b has sillimanite-biotite connected by tie-lines. The reaction which describes the change between these two figures: Biotite + Sillimanite = Cordierite + Almandine Garnet, is shown in Fig. 18 by univariant reaction line 2. The assemblage represented by Fig. 17a is stable under P-T conditions to the right of line 2, and that of Fig. 17b to the left of line 2.

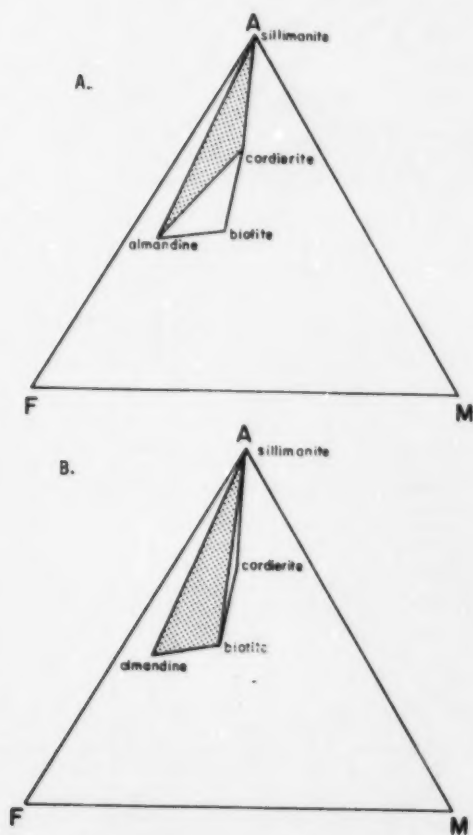


FIGURE 17a and 17b: AFM projections (after Reinhardt, 1968). See Figure 13 for legend. Stippled areas contain bulk-chemistry plots of pelites.

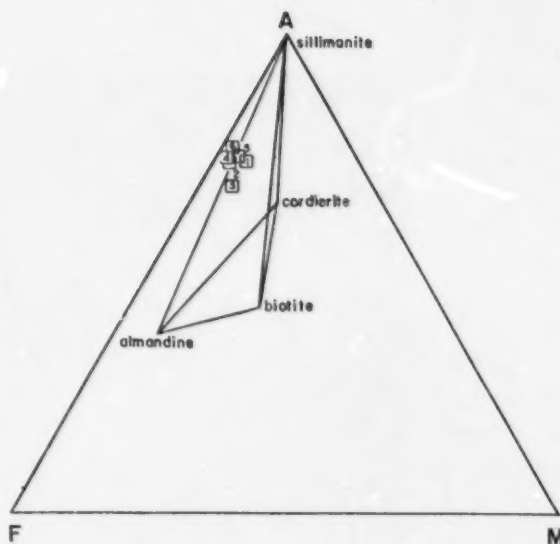


FIGURE 18: AFM projection (after Reinhardt, 1968). See Fig. 12 for legend. Mineral compositions are from unpublished data, Rutledge area metapelites.

Because of the bulk compositions of the rocks analysed, I propose that the stippled areas of Fig. 17a and 17b connect minerals of theoretically acceptable equilibrium assemblages in these rocks. Almandine garnet and cordierite are seen to coexist in both the Rutledge Lake Complex and the Mama Moose Complex sediments (except in the most northerly pod of the Mama Moose Complex); at least some of the biotite seen appears to be relict. Thus Fig. 17a appears to be more applicable to most of the rocks studied. In samples from the northernmost pod of the Mama Moose Complex, biotite does not coexist with any of the other phases shown on these diagrams, almandine garnet is not present, and sillimanite-cordierite occur together. Since the bulk compositions of these rocks do not lie on the sillimanite-cordierite tie line(s), no satisfactory explanation can be given for this in terms of the diagrams discussed above.

The coexistence of cordierite-garnet-sillimanite-quartz in both pelitic units confines the conditions of metamorphism to the area of P-T space spanned by the bivarient reaction: 3 Cordierite = 4 Sillimanite + 2 Garnet + 5 Quartz (reaction lines 3, Fig. 18). Analysis of the Fe:Mg ratio in the cordierite can be done to narrow this field down to a univariant reaction line. Analyses of Fe:Mg ratios in both cordierite and garnet can be used in several cordierite-garnet Fe-Mg exchange geothermometers/geobarometers for further estimates of P-T conditions (e.g., Lee and Holdaway, 1977; Currie, 1971).

The reaction of staurolite to produce garnet + hercynite spinel + sillimanite + quartz, a common assemblage in the Rutledge Lake Complex metapelites, is shown in Fig. 18 by univariant reaction line 4. Interactions between minerals of this assemblage and cordierite have been modelled by Schreyer (1976) within the system $\text{FeO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O}$, at a particular range of oxygen fugacities. Some of the results are shown by univariant reaction lines 5, 6, and 7 of Fig. 18.

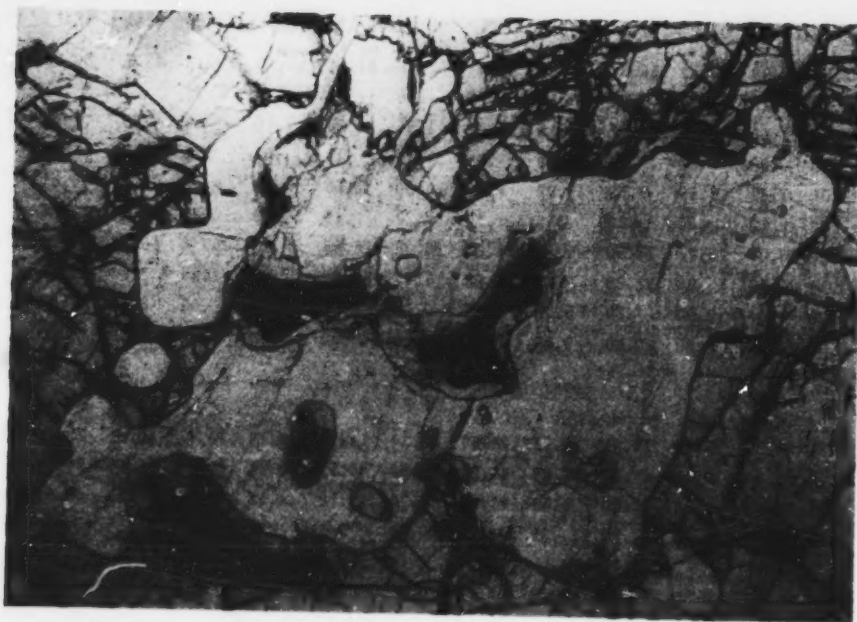


FIGURE 20: Photomicrograph of metasediment showing reaction between hercynite spinel (dark) and quartz (light, low relief) to produce sillimanite (fibrous material in garnet) and almandine garnet. Photo 2 mm along base.

A beautiful example of reaction 6 taking place is seen in a thin section (Fig. 20). Hercynite spinel reacts with quartz to produce a rim of sillimanite. The surrounding high-relief mineral is garnet with small fibers of sillimanite inclusions. The balanced reaction suggested here is: 3 Hercynite Spinel + 5 Quartz = Almandine Garnet + 2 Sillimanite. The theoretical phase relations above and below this univariant reaction line are shown in the Silica-Iron-Aluminum triangles of Fig. 19a and 19b respectively. Bearing in mind that these diagrams are applicable for an experimental system containing the Fe end-members of the mineral species present, the bulk chemistries plotted show a theoretical equilibrium assemblage of either Quartz-Almandine-Sillimanite, or Quartz-Hercynite-Sillimanite. Assuming equilibrium conditions have been reached, oxygen fugacity is of the right magnitude, and that the effects of other mineral phases present are minimal at the transition point between the two diagrams, equilibrium between all four phases is theoretically possible. The pressure-temperature conditions of metamorphism for this rock would therefore be on line 6, Fig. 18.

In one sample, hercynite spinel, sillimanite, cordierite, and almandine garnet are all intimately related (Fig. 21). The reaction: Cordierite + Hercynite Spinel = Almandine Garnet + Sillimanite, shown as univariant reaction line 7 (Fig. 18), may be applicable. Microprobe analyses of cordierites from several locales in the Rutledge Lake area pelitic sediments give Fe/Fe+Mg ratios of between 0.3 and 0.5, corresponding to possible pressures ranging from 3.5 to 5 Kb (Fig. 18).

Microprobe analyses of two-pyroxene pairs (OPX-CPX) on tholeiitic to komatiitic, mafic to ultramafic rocks from the Rutledge Lake area are plotted on the Pyroxene Quadrilateral in Fig. 22. These are applied

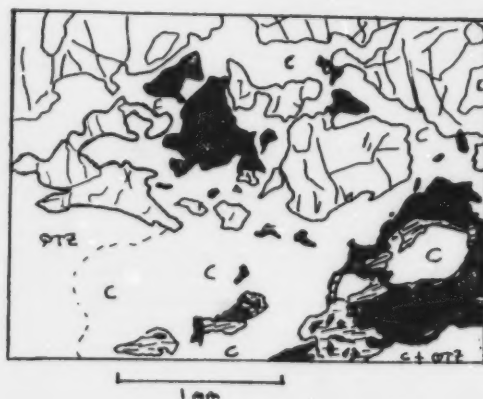


FIGURE 21: Sketch of metapelite sample 58 showing phase relations between hercynite spinel (black), sillimanite (fibrous), cordierite (C), and almandine garnet (high-relief, fractured mineral), quartz also present. PPL; scale bar marked.

in Table 2 to the exchange reaction and transfer reaction geothermometers of Kretz (1981). The results show temperatures ranging from 860° to 1030°C. Since total iron as FeO was used in these calculations instead of just ferrous iron, these temperatures may be too high by possibly 100°C (R. Kretz, pers. comm.). An average of the temperature calculations made is 960°C ± 60°C. It is important to say here that all pyroxenes analyzed are

TABLE 2: TWO-PYROXENE GEOTHERMOMETRY

OXIDES	OPX1	CPX1	OPX2	CPX2	OPX1	CPX1	OPX2	CPX2	OPX1	CPX1	OPX1	CPX1	OPX2	CPX2
SiO ₂	53.40	52.49	53.13	54.81	53.59	52.43	52.07	52.86	53.10	52.18	48.77	50.83	49.19	50.12
TiO ₂	.12	.12	.16	.01	.12	.08	.06	.05	.14	.17	.22	.22	.17	.22
Al ₂ O ₃	.96	1.65	1.06	1.67	1.16	1.69	1.36	1.54	1.10	2.09	1.07	2.40	.99	2.19
Cr ₂ O ₃	.20	.30	.22	.23	.10	.21	.15	.19	.08	.18	.08	.07	.04	.07
FeO	19.04	7.37	18.59	7.59	19.97	7.68	18.60	7.65	18.49	8.10	31.51	15.61	30.93	15.58
MnO	.32	.18	.35	.15	.36	.15	.37	.19	.59	.29	.54	.26	.50	.23
MgO	26.24	16.00	26.27	16.90	25.93	15.83	25.36	15.95	26.00	15.32	16.33	11.84	16.54	11.72
CaO	1.34	21.13	.80	22.03	.74	20.98	1.42	21.69	.61	21.85	1.04	19.60	.87	19.40
Na ₂ O	0.00	.07	0.00	.39	0.00	.40	0.00	0.00	0.00	.45	0.00	.33	0.00	.05
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00

OXIDE
PROPORTIONS

Fe	28.20	11.71	27.99	11.51	29.75	12.24	28.34	11.98	28.18	12.77	50.86	25.25	50.26	25.40
Mg	69.25	45.30	70.48	45.68	68.84	44.94	68.88	44.51	70.64	43.07	46.99	34.14	47.92	34.06
Ca	2.55	42.99	1.54	42.80	1.41	42.82	2.78	43.50	1.18	44.15	2.15	40.61	1.82	40.54
X*	0.29	.20	.28	.20	.30	.21	.29	.21	.28	.23	.51	.42	.51	.42
K _D **		1.63		1.58		1.61		1.54		1.30		1.44		1.44
[Ca]***		.43		.43		.43		.44		.44		.41		.41
T ₁ ^{°C}		860		920		880		930		930		1030		1030
T ₂ ^{°C}		1020		1030		1020		970		910		950		950
T _{avg} ^{°C}		940		975		950		950		920		990		990

* $X = \text{Fe}^{2+}/(\text{Mg} + \text{Fe}^{2+})$ Exchange Reaction: $T_1(^{\circ}\text{C}) = (1130/(\ln K_D + 0.505)) - 273$

$$K_D = \frac{X^{\text{OPX}}}{1-X^{\text{OPX}}} \times \frac{1-X^{\text{CPX}}}{X^{\text{CPX}}}$$

Transfer Reaction: $T_2(^{\circ}\text{C}) = 1000/(0.054 + 0.608^{\text{CPX}} - 0.304 \ln(1 - 2[\text{Ca}]^{\text{CPX}})) - 273$ *** $[\text{Ca}] = \text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}^{2+})$

(Geothermometers from Kretz, 1981)

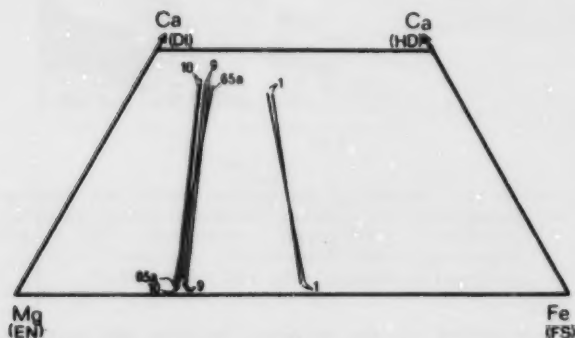


FIGURE 22: Locations of pyroxene compositions from microprobe analyses shown: CPI plots on high-Ca side of solvus; OPX plots on low-Ca side of solvus. Pairs used for geothermometry connected by tie-lines. Sample numbers follow scheme outlined in TABLE 1, mafic-ultramafic rocks (except for 85a, from the Nama Noose Complex).

believed from textural considerations to be metamorphic (they are xenomorphic, and commonly form triple junctions with neighboring crystals); larger, possibly relict igneous grains were avoided. Temperatures calculated are therefore not believed to record earlier magmatic conditions.

From consideration of the petrological phase relations in the metasediments and examination of the petrogenetic grid (Fig. 18) onto which the P-T conditions of theoretical relations between these phases are plotted, it is reasonable to estimate the pressure of metamorphism at between 3.5 and 5 Kb., corresponding to a depth in the earth's crust of about 10-15 km. Two-pyroxene thermometry points to metamorphic temperatures of 900°C or more, well into the granulite facies and above the solidus of a fairly dry ($P \text{ H}_2\text{O} = 0.1P \text{ Total}$) granite (Currie, 1971) (curve 8, Fig. 18). This is a high-T, low-P type of metamorphism, most likely a result of the intrusion of granitic magma which formed the plutons in the west, central, and east Rutledge Lake area. It is important to note that both the tholeiitic and komatiitic rocks record similarly high temperatures of metamorphism, evidence that both were emplaced before the intrusion of the granites.

The source of heat which caused melting and emplacement of the granites may have contributed to the metamorphism of the country rocks by causing upward displacement of the regional isograds (thermal doming), but this is entirely speculative.

One important observation is that the charnockitic gneiss (Unit 3) is a metamorphosed granite (showing good crystalloblastic texture) which must have been intruded early with respect to the other granites so that it could be re-heated to high-grade (hypersthene-forming) temperatures.

From descriptions of similar granulite-facies rocks to the south of the Rutledge Lake area, shown by Fraser (1978) to be Archean, the age of the granulite metamorphism, and therefore, by this model, the age of emplacement of the granites, is assumed to be Archean, possibly pre-Kenoran.

Later amphibolitization perhaps associated with mylonitization of the rocks and formation of the N-S trending shear zone which Rutledge Lake lies along, and greenschist metamorphism confined to northeast-trending minor shear zones, are not as apparent as the influences of the granulite-facies metamorphism. Growth of lower-grade alteration phases (e.g. muscovite, chlorite), existence of localized amphibolite bodies (not sampled), the presence of ribbons of strained quartz grains and other, megascopic structural features are perhaps the only evidence of later, lower-grade events seen in the Rutledge Lake rocks.

SUMMARY AND CONCLUSIONS

The rock units distinguishable in the field at Rutledge Lake are Western, Central (Rutledge Lake Complex), and Eastern Granites, a Charnockitic Gneiss, the Rutledge Lake and Mama Moose Complex Metasediments, and some Mafic to Ultramafic rocks. Upon chemical examination of examples of each of these units (Appendix 2), the following conclusions are drawn:

The granites are calc-alkaline intrusions. Their genetic relationship to each other cannot be postulated on the data available. Some of the granitic bodies have inclusions of aluminosilicate phases as xenocrysts and metasedimentary blocks as xenoliths, however, chemically they appear more to be mutually related than they appear related to the sediments which they intrude. This chemistry, appears to rule out the possibility that granites originated by mass-melting of the sediments. The two are concluded to be related by assimilation of the sediments by the granites.

The pelitic sediments have major-element bulk chemistries comparable to Henderson's (1972) compilation of Slave Province mudstones and have the appearance, where relatively undeformed, reminiscent of water-lain, mass-flow (turbidite) deposits. High-Mg can be explained as resulting from source-rock chemistry. Provenance of these sediments (whether from the Slave craton or another source) is not known. High-potassium in the sediments may be an effect of contamination from the intrusion of the granites and metamorphism of the sediments.

The mafic to ultramafic rocks can be classified on the basis of chemistry alone as either tholeiites or basaltic komatiites. They are not spatially or morphologically distinguishable from one another. High potassium and chromium are characteristic of the basaltic komatiites; the tholeiites show high-titanium affinities, possibly indicating an inter-

continental origin. High-potassium in the basaltic komatiites is attributed to contamination by the intruding granitic bodies and metamorphism of the area. Low-potassium in the tholeiites suggests that they intruded post-solidification of the granites, however since the tholeiites were metamorphosed to temperatures higher than the maximum melting temperature of the granite, this genetic interpretation is unlikely.

Perhaps the host-rock chemistry of the tholeiites, under thermodynamic conditions endured during metamorphism, was not correct to promote the nucleation or growth of potassium-bearing phases and hence potassium was not fixed in the rock, though it was in abundant supply. It is also possible that the tholeiitic rocks, during the movement of potassium throughout the area, had crystallized too completely to allow diffusion of potassium-bearing fluids through them.

The origin of the komatiitic rocks is not apparent, it is not known whether they are intrusive or extrusive. High Cu and Ni assays in showings which are intimately associated with komatiitic bodies make these rocks of special interest. A rigorous study of the economic geology of Rutledge Lake could shed further light on this association.

For the purpose of prospecting for Cu-Ni sulphides in the Rutledge Lake area, it is advisable to look specifically for mica-bearing mafic to ultramafic rocks and try to determine their spatial associations and geophysical signature.

Phase relations observed in the meta-pelites suggest a high-T, low-P granulite-facies metamorphism for the area. Two-pyroxene geothermometry done on several tholeiitic and komatiitic rocks from the area, combined with the petrographic observations, give a pressure estimate of about 3.5 to 5 Kb, and a temperature estimate of $960^{\circ} \pm 60^{\circ}\text{C}$. This is above the maximum melting temperature of all but the very driest granite. This type of metamorphism can be explained as resulting from the intrusion of the granites into the previously existing sediments and mafic to ultramafic rocks. These former country rocks now exist as rafted tracts, blocks, xenoliths and xenocrysts in the granites, all small enough to get heated up thoroughly by the intrusion of the magmas. These conclusions agree with those of Fraser (1978) on the granulite metamorphism found in the NW Churchill Province.

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APPENDIX 1: Thin section mineral log.

SAMPLE	Q	Pl	K-f	Si	C	G	Sp	Op	Bio	OPX	CPX	Se	Chl	Ap	Zir	Ep	UNIT
142 1a	X			X				X				X	X				NM
143b	X		X	X	X			X				X	X				NM
180		X						X		X	X						NM*
266b		X						X	X	X							NM*
266	X	X	X	X	X			X									NM
264	X		X					X	X								NM
270b	X	X						X	X	X	X			X			NM*
62b	X	X	X	X	X			X					X				NM
65a	X	X						X		X	X						NM*
66	X	X		X				X				X	X				NM
94a	X		X	X	X			X	X			X			X		NM
93c	X		X	X	X	X	X	X	X			X			X		NM
93a	X			X	X	X		X	X			X	X		X		NM
93b	X		X	X	X	X		X	X			X	X		X		NM
257b	X		X					X	X			X		X		X	EG
61a	X	X	X					X					X			X	EG
230b	X			X		X		X				X					RL
218	X	X	X	X		X		X				X					RL
201a	X	X		X		X		X	X			X					RL
211	X		X	X		X		X	X			X					RL
207	X			X		X	X	X				X	X				RL
197a							X	X		X, Olivine							RL*
245	X	X	X	X		X	X	X	X			X	X				RL
155b	X	X	X	X	X	X	X	X				X					RL
155c	X	X		X					X			X					RL
155d	X	X		X		X	X	X				X					RL
155a	X	X			X	X		X	X			X					RL
148	X		X	X		X		X	X			X					RL
164c	X	X				X		X	X			X					RL
171a	X	X	X	X		X	X	X	X			X					RL
139 1	X	X		X		X	X	X	X			X					RL
146b	X					X		X	X			X					RL
125a	X	X		X		X		X	X			X					RL
99			X	X	X			X	X			X	X		X		RL
100	X		X	X	X	X	X	X	X			X	X				RL
274a		X						X	X	X	X	X					RL*
80b	X		X	X		X	X	X		X	X	X				X	RL
73a	X			X	X			X	X			X					RL
74	X			X		X		X	X			X				X	RL
46a	X			X				X				X	X		X	X	RL
58	X		X	X	X	X	X	X	X			X			X		RL
166	X				X	X		X	X			X					RL
53a	X			X				X				X				X	RL
169	X	X	X					X	X								CG
256b	X	X						X			X						CG
149c	X		X	X		X	X	X				X				X	CG
162c	X	X	X					X				X					CG
131a	X	X	X						X	X							CG
20a	X	X	X					X				X	X				CG
196	X		X		X	X	X	X				X					WG

Unit designations defined in Table 1.

Asteriz specifies mafic to ultramafic rock enclosed in unit marked.

Samples listed North to South-top to bottom for NM and RL units.

Q = Quartz
Pl = Plagioclase
K-f = K-feldspar
Si = Sillimanite
C = Cordierite
G = Garnet
Sp = Spinel
Op = Opaque minerals

OPX = Orthopyroxene
CPX = Clinopyroxene
Se = Sericite
Chl = Chlorite
Ap = Apatite
Zir = Zircon
Epi = Epidote

APPENDIX 2: Whole-rock analyses. All from XRF except those marked "*" from DCP.

	1	2	3	4	5	6	7	8	9	10	11	12
	180	86c	238b	279b	253	76b	174	127c	270b	266b	105a	197
%SiO ₂	49.18	42.94	48.34	45.73	48.76	52.48	50.27	51.17	48.92	51.25	47.2	46.2
Al ₂ O ₃	13.33	14.02	14.66	13.23	13.44	16.13	13.19	12.33	10.27	9.58	8.1	6.5
Fe ₂ O ₃ T	18.34	17.87	14.86	15.76	14.41	10.54	13.24	10.28	11.26	12.28	2.5 - 2.8 Fe ₂ O ₃ 9.4 - 9.6-FeO	
MgO	6.37	6.42	6.65	7.73	7.82	8.25	8.40	11.84	15.49	16.80	20.4	22.5
CaO	8.16	10.57	10.26	14.13	11.56	6.18	10.36	3.59	6.99	6.22	5.56	7.78
Na ₂ O	2.35	2.33	2.29	2.13	2.14	3.35	2.46	0.48	1.46	1.56	00	0.2
K ₂ O	0.55	0.86	0.61	0.44	0.54	1.48	0.68	5.28	2.75	1.79	2.76	0.46
TiO ₂	2.49	3.19	1.43	1.27	1.38	1.01	1.34	0.87	0.72	0.72	0.75	0.55
P ₂ O ₅	0.08	0.30	00	00	00	0.44	00	0.40	0.20	0.25	0.22	0.09
MnO	0.26	0.23	0.23	0.25	0.22	0.16	0.21	0.12	0.18	0.20	0.20	0.23
S	0.02	0.09	0.03	00	0.02	0.02	0.04	0.01	0.01	00	0.21	0.20
PPM Ba	28	357	124	59	90	1147	107	967	1162	1122	640	99
Cr	131	70	109	203	128	317	311	820	1093	1250	1500	2000
Zr	97	191	65	40	66	172	170	149	99	89	.006%	00
Sr	166	433	131	124	141	434	155	78	166	325	55	110
Rb	6	8	44	4	12	62	00	295	117	113	.012%	00
Y	25	36	26	34	23	22	22	8	23	22	42	47
Nb	12	45	4	00	3	8	4	9	7	10	na	na
Zn	144	147	121	97	102	114	106	109	98	121	.010%	.009%
Ni	69	48	61	100	108	99	114	156	353	388	510	610
TOTAL	101.2	99.01	99.45	100.8	100.4	100.3	100.3	96.69	98.67	101.10	99.6	99.7
Y*	na	na	13	16	10	9	11	6	na	10	10	8
Cr*	199	na	na	na	na	na	na	na	1390	na	na	na
Zn*	185	na	na	na	na	na	na	na	119	na	na	na
Cu*	27	na	na	na	na	na	na	na	41	na	na	na

	1	2	3	4	5		1	2	3	4	5	6
	257b	114	169	61a	196		53a	93b	171a	62b	142 1a	143b
%SiO ₂	66.56	69.74	70.55	71.73	73.39		59.40	60.74	63.36	65.41	71.15	71.71
Al ₂ O ₃	15.67	15.59	14.81	14.19	14.55		22.79	24.47	18.05	19.38	15.39	13.41
Fe ₂ O ₃ T	3.76	2.06	3.04	1.84	1.26		6.90	8.52	8.40	6.76	5.26	5.11
MgO	1.28	0.56	0.92	0.38	0.20		2.00	1.43	2.26	1.07	1.09	0.58
CaO	3.18	1.06	2.95	1.42	0.65		0.14	0.09	0.82	0.50	0.10	0.59
Na ₂ O	4.17	2.36	4.32	3.84	2.91		0.36	0.43	1.65	0.94	0.56	1.49
K ₂ O	2.97	6.48	2.96	4.49	5.83		4.02	4.07	3.66	2.96	3.53	4.42
TiO ₂	0.42	0.27	0.36	0.18	0.09		1.01	1.04	0.78	0.93	0.72	0.57
P ₂ O ₅	0.13	0.12	0.08	0.02	0.14		0.02	0.04	0.05	0.03	0.03	0.02
MnO	0.08	0.02	0.06	0.03	0.01		0.02	0.05	0.11	0.06	0.03	0.04
S	0.01	0.01	0.01	0.02	0.00		0.07	0.00	0.03	0.00	0.01	0.00
PPM Ba	2123	704	925	1009	262		702	796	770	550	601	829
Cr	46	50	36	32	26		228	158	113	130	97	74
Zr	145	106	120	129	34		132	151	196	234	167	299
Sr	601	173	230	191	67		40	96	127	48	59	78
Rb	43	225	81	94	269		184	157	124	115	145	176
Y	11	7	11	7	10		10	19	52	19	12	11
Nb	6	6	7	5	7		13	19	11	21	15	17
Zn	57	70	40	14	36		117	125	109	92	69	80
Ni	23	26	12	21	15		88	46	54	42	43	31
TOTAL	98.60	98.43	100.25	98.32	99.13		96.90	101.1	99.37	98.17	98.01	98.12
Y*	na	na	na	na	na		na	na	na	na	na	na
Cr*	91	60	57	32	24		250	190	136	146	119	774
Zn*	83	60	60	25	36		133	143	127	106	75	86
Cu*	00	00	1	44	00		103	00	00	00	00	00

STRATIGRAPHIC EVIDENCE FOR THE
EXISTENCE OF A GLACIAL LAKE IN THE
RICHARDSON AND RAE RIVER BASINS, N.W.T.

by

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ABSTRACT

Subglacial tunnels evidenced by eskers or esker fans were formed at the ice front where sediments flowing into the proglacial lake provided a suitable environment for subaqueous outwash deposits. A distinctive facies change in the downcurrent direction is thought to represent a progressive decrease in flow velocity. For this reason, the esker to varve transition is characterized by the following bedding structures: a) planar and cross-bedded proximal gravels deposited near the mouth of an ice tunnel, b) massive to bedded, poorly sorted medium to coarse sand, c) fine sand and silt exhibiting type A and B cross-laminations and sinusoidal laminations, and d) glaciolacustrine rhythmites consisting of sandy silt and clay couplets representing the finer fractions which were carried out into the lake by turbidity currents. Within the sedimentary succession studied, the glacial and proglacial environments are represented by an upward-fining trend, succeeded by an upward-coarsening trend corresponding to the regressive marine sequence.

RESUME

L'apport de sédiments dans un lac proglaciaire à partir d'un tunnel subglaciaire crée des conditions favorables à la formation de cônes sub-aquatiques. Un changement distinct de faciès observé à partir de la source représente une baisse progressive de la vitesse du courant. Par conséquent, la transition de l'esker aux varves est caractérisée par les structures suivantes: a) des graviers à stratification plane et interstratifiées qui sont déposés à l'embouchure du tunnel sub-glaciaire, b) des sables moyens à grossiers, mal triés et d'apparence massive à stratifiée, c) sable fin et silt à stratifications entrecroisées sinusoidales et de type A et B, et d) rhythmites glacio-lacustres de silt sableux et d'argile qui représentent les sédiments fins transportés dans le lac par les courants de turbidité. Dans les séquences sédimentaires étudiées, les environnements glaciaires et proglaciaires sont représentés par une séquence de grosseur de sédiment décroissante vers le sommet, suivie d'une séquence inverse qui correspond à la séquence de régression marine.

INTRODUCTION

Numerous stratigraphic sections, exposed primarily along the Richardson River (Fig. 1), have yielded valuable evidence concerning the history of the deglaciation in the Richardson River basin. A sequence of glacial, glaciolacustrine, glaciolacustrine and post-glacial marine deposits indicate a relatively clear succession of environments of deposition, commencing in the earliest stages of deglaciation and progressing through to the present day.

In this study, stratigraphic and sedimentological data are used to describe the various sedimentary environments that cannot be inferred by the mapping of surficial deposits alone. The study of such depositional sequences provides more detailed information that may serve to support and complete the reconstruction of the deglaciation history of an area.

PREVIOUS STUDIES

Until recently, only general observations con-

cerning the Quaternary geology of the area west of Coronation Gulf (Fig. 1) have been reported. The main interests of the early explorers and other geologists were in large part unrelated to the Quaternary geology (Franklin, 1828; O'Neill, 1924; Baragar and Donaldson, 1973).

However, Craig (1960) mapped the ice-flow indicators in the area and the extent of the marine limit, estimated at 146 m a.s.l. From his work it was also concluded that the ice retreated easterly over this area. More recently, a detailed study was undertaken by St-Onge (1980) in the Coppermine River area (Fig. 1). Systematic mapping of the surficial deposits helped define the pattern of deglaciation in the area. Following a stratigraphic study, St-Onge and Bruneau (1982) established the marine limit at 170 m a.s.l. at the mouth of the Coppermine River. To the west, further mapping of the surficial deposits in the Richardson River basin (Mercier, 1985) revealed the existence of a glacial lake (Mercier, 1984). This glacial lake was limited to the depression of the Rae and Richardson River basins, west of Coronation Gulf by the retreating ice front.

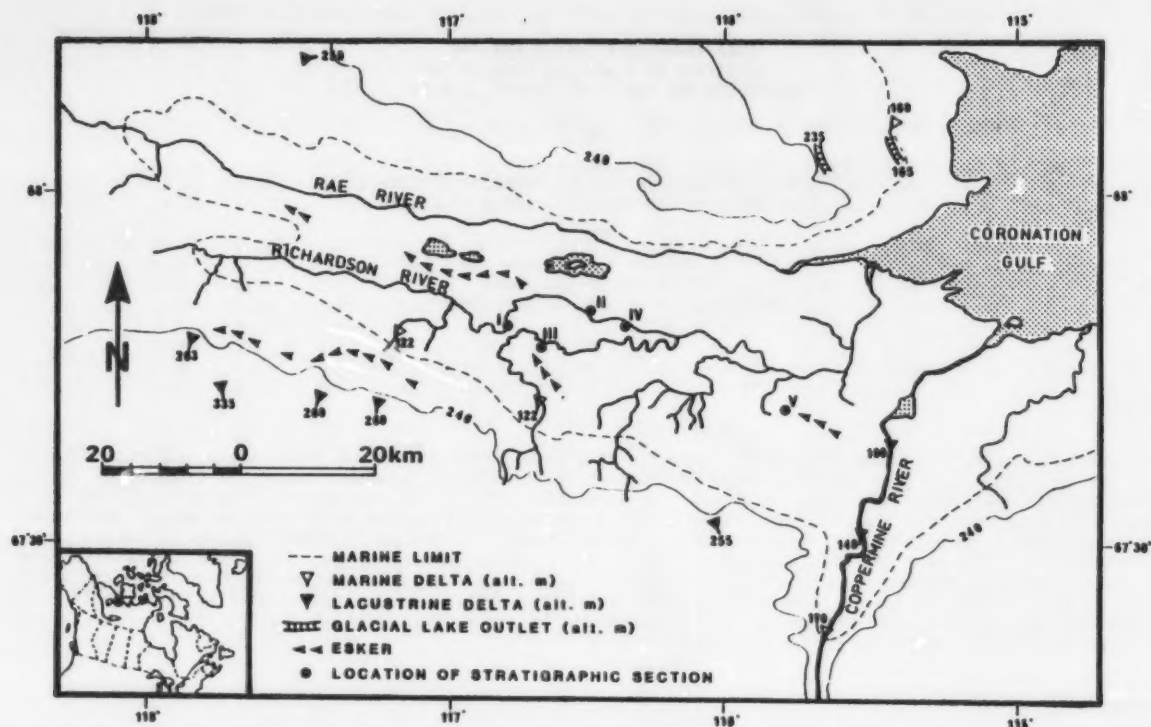


FIGURE 1: Location map and pertinent topographic features.

REGIONAL SETTING

Pre-Quaternary geology

Most of the study area (Fig. 1) is underlain by a volcano-sedimentary sequence of Proterozoic age. The volcanic formations of the Coppermine River Group dip northward at angles of 5 to 10 degrees and are overlain by the Rae Group, which forms a gently northward-tilted homoclinal sequence of sedimentary rocks (Baragar and Donaldson, 1973). Throughout most of the area the bedrock geology is obscured by thick surficial deposits.

Physiography

The Richardson River basin covers more than 5,000 km² and can be generally subdivided into two distinct physiographic zones. The southern part of the basin is characterized by a relatively unbroken plateau whose altitude decreases eastward. To the north, the relatively flat lowlands are filled by Quaternary sediments of glacial, glaciolacustrine and marine origin. The surface is occasionally punctuated by outcrops.

QUATERNARY SEDIMENTARY FACIES

Several different lithological units have been identified within the stratigraphic sections studied in the Richardson River basin (Fig. 1). In order to facilitate their description, they have been

classified as facies on the basis of the distinguishing characteristics of grain size and primary sedimentary structures. These facies not only reflect a wide range of energy levels, from upper to lower flow regimes, but are also representative of a wide spectrum of environments of deposition. While most facies are relatively widespread and found throughout the basin, a few are restricted to specific sites. The following is a brief description of each facies and its associated properties.

Diamicton Facies (Facies 1 & 2)

Facies 1 is a poorly sorted diamicton; the dark grey-brown matrix is composed primarily of clay and contains lesser amounts of silt, sand and gravel. The larger matrix-supported clasts, which consist of various lithologies, are rounded to angular and up to 1.0 m in diameter. This facies is represented by two distinct massive beds restricted to Section IV (Fig. 2) and separated by a bed of Facies 2. The upper bed is 1.5 m thick; the lower bed is 6.5 m thick and rests directly on Proterozoic sandstone. The upper and lower contacts are sharp and planar to irregular.

Facies 2 consists of a grayish silt-rich diamicton. In addition to silt, it is composed of sand, gravel and boulders up to 1.5 m in diameter. Its clasts are also rounded to angular and of variable composition. It is found at two sites, Section II and Section IV (Fig. 2). At the former, it appears as a massive, 4.0-m-thick bed with sharp

upper and lower boundaries that appear to be erosional. Within Section IV, it occurs as two beds, above and within Facies 1. Occasionally, poorly developed bedding is observed, with the thickness of beds ranging from less than 0.5 m to 5.0 m. The upper part of the unit interbedded within Facies 1 is characterized by small lenses of sandy gravels that do not exceed 75 cm in length.

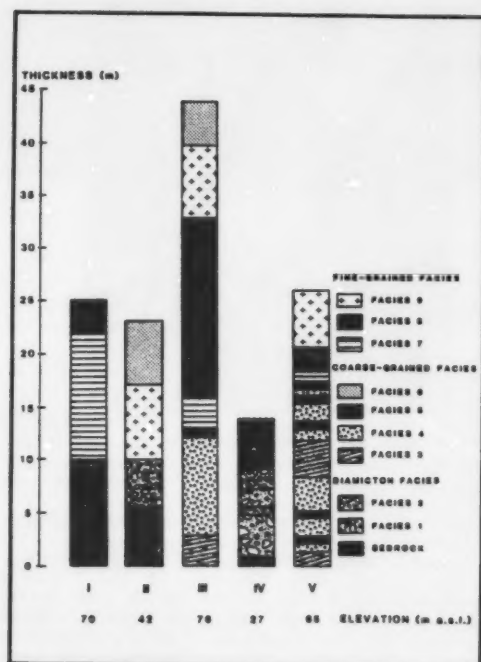


FIGURE 2: Sedimentary succession of facies in sections I to V.

Coarse-grained Facies (Facies 3 to 6)

Facies 3 is a sandy gravel that contains well-rounded cobbles up to 15 cm in diameter. It shows poor sorting as grain sizes vary from fine sand through to pebbles and cobbles. Facies 3 is composed of massive to well-bedded, 20-cm to 40-cm-thick strata that commonly dip 18° to 21° northwesterly. It is restricted to the base of Sections III and V (Fig. 2), where the observed maximum thickness is 3.0 m. Whilst upper contacts are gradational with Facies 4, the nature of the lower contact with underlying sediments could not be ascertained.

Facies 4 is a poorly sorted dark-gray sand with grain sizes in the medium to coarse range and rare granules. It is generally massive, but locally may exhibit horizontal laminations and vertically stacked small-scale cross-bedded cosets. Thickness of the cosets varies from 20 cm to 50 cm, and they are separated from one another by planar, erosional contacts. The angle of dip varies from 18° to 34° in a number of directions, ranging from southwest to northwest. This predominantly homogeneous unit is extremely variable in thickness, i.e., from 0.4 m to greater than 12.5 m (Section V, Fig. 2).

Facies 5 is a fine-grained to medium-grained grayish sand consisting of cross-laminated, small-scale cosets. With reference to the cross-laminations, the classification of Jopling and Walker (1968) has been adopted in view of the wide variation of laminated sedimentary structures.

The least common bedform is entirely composed of climbing sets of well-defined lee-side laminae with no preservation of stoss-side laminae (Fig. 4). These are type A cosets, which are developed in the coarse sand fraction and have low angles of climb. Those observed were found to be not more than 10 cm thick.

The second type of ripple-drift cross-lamination, type B, is composed of climbing sets of lee-side laminae with complete preservation of stoss-side laminae (Fig. 4). These laminae are continuous from one asymmetrical ripple to another. Each set, varying from 1.0 cm to 5.0 cm in thickness, may occasionally be overlain by a draped layer consisting of parallel clay laminations ranging in thickness from 0.2 cm to 0.5 cm. Sequences of uninterrupted cosets of type B cross-laminations do not exceed 1.0 m in thickness, and they show evidence of higher angles of climb than those of type A. Closer examination of some of these sets reveals that they are composed of alternating light and dark parallel laminae. For the most part, the crests and troughs of ripples are gently rounded, although sharp-crested peaks sometimes occur at the base of the individual sets. Palaeocurrent measurements of type B ripple-drift cross-laminations indicate variable flow directions from 270° through to 315° .

Undulating laminae or sinusoidal ripple laminations are also present, but only as a minor component. They consist of superimposed sinusoidal laminations, which are also laterally continuous. Both lee-side and stoss-side of the ripples are equal in thickness.

Facies 6 ranges from a poorly sorted sand, varying in grain size from silt to granules, to rhythmites. The thickness of this facies is also variable, from 1.0 m to 6.0 m (Section II, Fig. 2). The lower parts of this unit are structureless, but change upwards to discontinuous laminations, small pebbly lenses and finally to alternating layers of silty sand and silty clay. The gray, silty sand beds average 10 cm thick and do not show any internal stratification or grading within the layers. The upper and lower boundaries are gradational with the silty clay bands, which are structureless and dark gray in color, range in thickness from 2.0 cm to 10.0 cm and increase in thickness towards the top.

Fine-grained Facies (Facies 7 to 9)

Facies 7 has grain sizes within the silt to fine sand range and possesses a homogeneous light gray appearance due to the dominance of silt-sized particles. Bedforms vary from massive, through horizontal finely laminated silts to sinusoidal laminations and type B ripple-drift cross-laminations. This sequence, whether complete or partial, forms individual beds 0.3 m to 2.5 m thick that are separated by 1.0-cm to 10.0-cm-thick red-brown clay layers. Restricted to the middle of Section I (Fig. 2), the type B cross-lamination sets of Facies 7 are 30 cm to 40 cm thick and are overlain and underlain by transitional undulating laminae. These ripples have sharper crests than those of Facies 5, although their angles of climb are quite similar. Palaeocurrent directions are towards 250 degrees. Neither draped clay laminations



FIGURE 3: Till of Facies I overlain by glaciolacustrine rhythmites of Facies 8. Scale is 45 cm.



FIGURE 4: Sequence of type A ripple-drift cross-laminations overlain by type B ripple-drift cross-laminations with clay-draped laminations. Scale is 23 cm.

nor alternating dark and light laminae were observed within the ripple cosets.

Facies 8 is a major component of most of the sections studied. This facies resembles varve-like couplets consisting of two texturally and genetically distinct layers: a light gray silt and a darker brownish-gray clay. The silt layers, being consistently thicker than the clays, are composed of discrete horizontal laminations of varying grain size and thickness, numbering up to 45 per layer (Fig. 5). These internal laminae are characteristic of the



FIGURE 5: Proximal glaciolacustrine rhythmites showing multilaminated nature of the summer layers. Scale is 22 cm.

silt layers within the basal couplets of sections; no such laminations were observed in the rhythmites stratigraphically higher. Occasionally sand partings are present within some of these layers, and more rarely couplets are separated by thin diamicton beds less than 20 cm thick. Although the thickness of the silt layers is relatively constant within a given vertical range, the samples measured within the basin varied from less than 1.0 cm to 40.0 cm thick at the base of sections (Fig. 6).

Within a couplet, the contact between silt and overlying clay is gradational. Boundaries separating clay layers from overlying silt layers are sharp, but irregular erosional contacts have also been observed in some localities.

The thinner darker clay layers range from a few mm thick in the upper rhythmites to 10.0 cm in basal couplets. They are predominantly massive, but may exhibit poorly defined internal laminations within the thicker clay layers.

Generally, there is an overall trend for the couplets to decrease in thickness upwards, although there are local variations. The rhythmites do not appear to thin or thicken laterally within a given section, although regionally the thickness of Facies 8 varies considerably, from approximately 2.0 m in the eastern part of the basin to at least 25.0 m in the central part.

Other features present in Facies 8 include intraformational crumplings and dropstones, attaining 1.5 m in diameter.

Facies 9 is composed of dark clay, which is for the most part massive and structureless. Upper contacts are gradational with overlying facies, whereas the lower boundaries are planar to sharp and

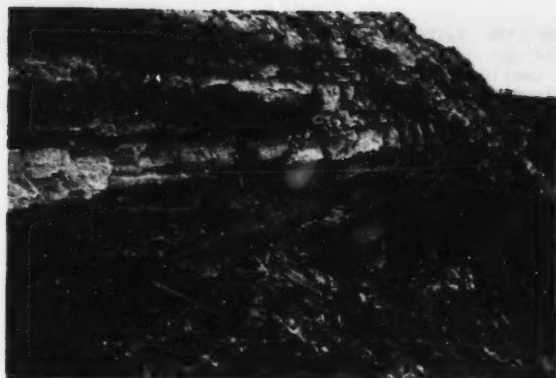


FIGURE 6: Thick sand-silt (light) and alternating clay (dark) intervals of Facies 8. Scale is 21 cm.

irregular (Mercier, 1984, Fig. 2). Its thickness varies from 5.0 m to 15.0 m in the eastern and central part of the basin and decreases to as little as 0.4 m further west.

FACIES RELATIONSHIPS

In order to provide a plausible interpretation of depositional processes and environmental conditions, it is necessary to define specific facies associations which group together different sedimentary facies that are environmentally or genetically related. Four such associations have been noted.

In the case of Section IV, the interfingering and interbedded nature of Facies 1 and 2 is indicative of an environment in which slight variations in deposition are occurring. As Facies 1 has been classified as till (see discussion), it can be inferred that Facies 2 is also of glacial origin in view of its similarities and relationship with Facies 1.

The sedimentary sequences of Facies 3, 4, and 5 (Sections III and V) represent an environment in which there is considerable fluctuation in the flow regime. These are all basically coarse-grained sediments characterized by the presence of Facies 3 at the base of sections, overlain by Facies 4 and is, in turn, followed by Facies 5 which appears to be a transitional phase between Facies 4 and 7, although it is also found interbedded with Facies 4. The vertical position of these three facies within the sections studied, as well as in several others not discussed in this paper, is consistently the same throughout the basin: they underlie thick successions of silt and silty-clay rhythmites, forming part of a general fining-upward sequence. It is probable that Facies 5 and 4 are the more distal equivalents of Facies 3, which represents a proximal deposit closer to its glacial source. The variable thickness of these and other facies can be the result of the effects of three factors: the intensity, the duration and the frequency of discharge cycles.

A third sedimentary association is that of Facies 7 and 8; the latter dominating the majority of the sections in the Richardson River basin. The silts of Facies 7 appear to be gradational from Facies 5 to 8 (Sections III and V). In this context, they are laterally extensive, occurring at the base of Facies

8 at many sites throughout the basin, although they are not volumetrically significant. An exception to this is found at Section I where Facies 7 is underlain and overlain by Facies 8. This poses a particular problem with respect to the interpretation of this sequence and will be discussed later.

Noticeable lateral variations in thickness of Facies 8 have been observed in the eastern and central parts of the basin. In general, the thickness of the facies is greatest in the northern areas of topographic lows, and thinner further east and west in regions representing topographic highs.

Facies 9 and 6 form the fourth facies association from which a coarsening-upward cycle has been identified. The massive clay unit abruptly overlies Facies 8 in Sections III and V and Facies 2 in Section II. It grades gradually into Facies 6, which is more common and extensive to the east. Other than in areas where surface erosion has taken place, these facies overlie all others, completing a diverse and complex sedimentary succession.

DISCUSSION

Evidence that a dammed glacial lake once occupied the depression formed by the basins of the Rae and Richardson Rivers (Fig. 1) was first discovered by Mercier (1984). It is thought that Glacial Lake Richardson evolved through four principal lake phases, each controlled by successively lower outlets. In the glacial context, the presence of this lake provides a framework around which different depositional environments developed.

The eastward-retreating ice front of a lobe of the Laurentide ice sheet dammed glacial meltwater during retreat of ice from the basin. Numerous eskers and glaciofluvial complexes whose predominant orientation is N.W.-S.E. (Fig. 1) were developed during this period (Mercier, 1985). Two of these ice-contact deposits are recorded at the base of Sections III and V (Facies 3, 4, and 5), and are believed to be subaqueous in origin. They consist of stratified sand and gravel removed from a glacier by meltwater streams and deposited in front of and beyond the glacial margin into a standing body of water (Rust and Romanelli, 1973). These subglacial eskers are readily visible throughout the basin as ridges of glaciofluvial sediment forming raised, sometimes sinuous landforms.

Some of these sediments are relatively coarse-grained, such as Facies 3, and for this reason must have accumulated close to their glacial source. Changes in current velocity are responsible for the deposition of Facies 4 and 5, whose diversity of sedimentary structures and grain sizes is the result of the variability of discharge, diurnally to seasonally (Saunderson, 1973), in a glaciofluvial environment.

Facies 1 and 2 are believed to be a second type of ice-contact deposit. These two interlayered diamicton facies were only observed within two stratigraphic sections (Fig. 2). The similarity of Facies 1 to surficial till within the Richardson River basin suggests that this facies may be a subglacial lodgement till deposited from actively moving ice. According to Boulton (1972), in some areas the retreat of a glacier is initially represented stratigraphically by till with minor stratified beds and lenses within it. The till tends to lie on smooth or striated bedrock.

The origin of Facies 2 is less well understood; it could represent a flow till (Boulton, 1972, p. 379), and local upward fining of the poorly stratified diamicton suggests that it was water-lain before another layer of till accumulated above it. Three possibilities have been envisaged to interpret the occurrence of Facies 2 in Section III. It may have originated as ice-rafted debris during calving of the ice front, or it could be the product of slumping of a semi-fluid till off the ice front or by direct glacial deposition during a minor glacial surge. An alternative interpretation is slumping or failure of deposits accumulated along the emerged lake shorelines (Gilbert, personal comm., 1984). Further field investigation is required to determine which is the most probable hypothesis.

Escher gravels and varves have commonly been found adjacent to each other (De Geer, 1940). These gravels change laterally into varve couplets, becoming finer-grained in a distal direction. In the Richardson River basin, a similar vertical sequence occurs in which a vertical equivalent to a single horizontal time-stratigraphic unit is deposited. The succession from Facies 3 through to Facies 7 can be explained by a gradual increase in the ratio of suspended load to bed load in the downcurrent direction (Saunderson, 1973). For this reason, coarse esker gravels (Facies 3) are deposited near the mouth of a subglacial tunnel and as the competency of the flow decreases in the distal direction, the sediments become progressively finer-grained. Saunderson (1973) states that this particular facies sequence is diagnostic of the transition that occurs between eskerine deposits and sediments that are deposited in relatively deep water. Stratigraphic and geomorphic evidence supports the hypothesis of a relatively deep (200 m) glacial lake in contact with the ice front during deposition of outwash deposits and proximal, as well as distal, lacustrine rhythmites.

The change from thick sandy-silt proximal rhythmites (Facies 5 and 7) to thinner distal lake varves (Facies 8) is a gradual one. This is accompanied by a decrease in grain size in the coarser summer layer and a change in the dominant sedimentary structures. Each varve couplet represents a seasonal variation in melt with the thick silts being deposited in summer by turbidity currents and the thin clays settling out of suspension during the winter. According to Shaw (1977), where the glacial retreat rate is high compared to the rate of sedimentation, varved silts and clays are expected to dominate the succession. This phenomenon is most noticeable in the central part of the basin where there is a 25-m sequence composed entirely of glaciolacustrine rhythmites.

The upward decrease in thickness of couplets is attributed to deposition in increasingly distal positions as the ice front retreated. This trend is characteristic of classical glacial lake sediments deposited during ice retreat (Shaw, 1977).

In Section I, the glaciolacustrine rhythmites (Facies 8) are gradually interrupted 10 m above the base of the section by increasing then decreasing thicknesses of sandy silt beds (Facies 7), alternating with a thinner clay layer identical to that of Facies 8. The presence of this facies interlayered with a facies representing a more distal, comparatively tranquil environment indicates an increase in the rate of sedimentation from suspension and/or increase in flow velocity similar to that which produced the proximal glaciolacustrine rhythmites of Facies 7 in Section III and V. A possible explanation

for the increase in the silt layer and change in sedimentary structures can be interpreted as a local record of change in the direction of meltwater discharge from a nearby glacial front. An alternative to this is a minor readvance allowing the sequence of subaqueous outwash facies to temporarily prograde upgradient further west into the basin. However, the lack of good exposures studied to date makes it impossible to correlate some of the stratigraphic units or to trace them laterally to their corresponding coarse-grained outwash deposits.

The rhythmites of Facies 8 have been interpreted by Dr. D.A. St-Onge (personal comm., 1984) as marine, rather than to glaciolacustrine in origin. This conclusion is based on two points: 1) the presence of up to 70 m of silty-clay rhythmites found in the Coppermine River valley (Fig. 1) below the marine limit (170 m a.s.l.), between 154 m a.s.l. and present-day sea level and 2) the apparent absence of rhythmites and silt and clay deposits between 260 m and 150 m a.s.l. in the Richardson River basin.

The fact that these rhythmites are found below 150 m a.s.l. (marine limit in the Richardson River basin) does not necessarily imply a marine origin and can easily be explained by two alternate interpretations: 1) the fluid behaviour of turbidity currents: the fines transported by these currents will tend to be deposited in the deepest part of the basin where they may accumulate in a relatively tranquil environment (Collinson and Thompson, 1984; Shillis, personal comm., 1985), and 2) the region lying between 260 m and 150 m a.s.l. was exposed to a high level of current action and subsequent wave action which largely inhibited the widespread deposition of silts and clays during the period of time this area was submerged by Glacial Lake Richardson.

Some deltas at altitudes of 330 m and 260 m a.s.l. (Fig. 1) show evidence of collapse structures and kettle lakes, indicating that they were ice-contact deposits built against or in proximity to the ice lobe which occupied the Richardson River basin. These early lake phases are associated with a marginal proglacial lake in which the 330 m (a.s.l.) deltas and one 260 m (a.s.l.) delta were formed. The remaining deltas at 260 m (a.s.l.) were built in the lake when the latter became progressively larger as ice continued to retreat eastward.

A 17-km-long glacioluvial esker-delta complex (Fig. 1) also developed during the early stages of ice retreat when Glacial Lake Richardson was still a marginal lake. It is probable that a substantial amount of the finer outwash sediments were transported through and out of this system by currents and carried away towards the northwestern part of the marginal lake, which drained through a channel further north (Mercier, 1984).

It is also conceivable that the oversteepening of delta fronts may have led to slumping of foresets resulting in grain flows which deposited some of the sand found below 260 m a.s.l. Erosion of fines deposited downslope could also have occurred during this process (Gilbert, personal comm., 1984; Shaw, 1977).

The effect of these high-energy environments would be to remove silts and clays, thus accounting for the absence of fines on some of the steeper and more elevated slopes of the basin between the altitudes of 260 m and 150 m a.s.l. It should also be pointed out that the sparseness of fine-grained lake-bottom sediments on the slopes of lake basins is not

uncommon and is known to occur both in numerous modern lacustrine environments as well as in Pleistocene glacial lakes (as seen in profile in Shilts, 1984; Shilts and Farrell, 1982).

The source of the sediments composing the rhythmites is also an essential factor to consider when determining the origin of these rhythmically bedded deposits. In view of the fact that this facies attains 25 m in thickness and is relatively widespread, such a significant amount of fines could only have originated principally from glacier ice. Although some of these silts and clays undoubtedly are of terrestrial origin, it is improbable that the rivers draining the basin of the Richardson River could have supplied such a large amount of sediment to the sea in order to form the rhythmites.

The upper reaches of the basin are underlain predominantly by bedrock and till veneer. There is a marked absence of any major river system and erosional features. The small number of marine deltas and their modest dimensions attest to the minor role rivers must have played as sediment-supplying agents.

The continuous eastward retreat of the ice front permitted lake levels to fall as lower glacial lake outlets were freed of ice. Shortly after Phase 4 (Mercier, 1984, Fig. 9), the ice front retreated still further east and eventually permitted the incursion of the sea to its maximum altitude of 170m a.s.l. (St-Onge and Bruneau, 1982), which marked the end of Glacial Lake Richardson. This relatively sudden and short-lived event is reflected in the stratigraphic record by an abrupt change in the mechanism of sedimentation, which therefore resulted subsequently in the deposition of two totally dissimilar sedimentary facies. This can be seen in Sections III and V where the massive marine clays of Facies 9 overlie the glaciolacustrine rhythmites of Facies 8. This event occurred prior to $10,300 \pm 240$ B.P. (GSC-3663), a radiometric date obtained from marine shells collected in a raised beach situated at an elevation of 122 m a.s.l.

Isostatic uplift caused the gradual regression of the sea to its present level. During this period, a coarsening-upward sequence developed due to the lowering of sea level, as evidenced in Section III in which Facies 9 grades vertically into Facies 6. A detailed stratigraphic and sedimentological study of the latter stages of marine regression within an estuarine environment by Kerr (1984) describes the transition from shallow marine waters through to shallow subtidal and intertidal environments, followed by a brief period of fluvial activity. This sedimentary succession of marine deposits is known to have occurred before $1,940 \pm 80$ B.P., (B.G.S.-912) as peat within the fluvial sediments at 25 m a.s.l. provide this age.

In addition to this, it is known that structureless, massive marine silts and clays (Facies 9) had been accumulating in the basin since at least 9,400 B.P. (preliminary date, GSC-3941) and continued to do so as a continuous process throughout the marine regression (Kerr, 1984) to the present. These massive fine-grained sediments possess abundant and highly varied marine micro- and macrofossil assemblages, whereas the rhythmites are devoid of fossils.

On the basis of these hydrological considerations, coupled with the sedimentological evidence supporting an obvious genetic association between glaciofluvial outwash deposits (Facies 3, 4 and 5) and Facies 7 and 8, it is the authors' belief that

these rhythmites are indeed entirely glaciolacustrine in origin.

CONCLUSION

During the easterly retreat of the ice sheet, Glacial Lake Richardson experienced a wide variety of rapidly changing environmental conditions due to the effects of an unstable ice front.

The depositional sequences studied record a series of distinct sedimentary facies characteristic of particular environments, each representing a different energy level and position within a well-defined succession.

The data collected from ice-contact glacial and glaciofluvial deposits have provided a better understanding of sedimentary processes in deglaciated regions. The analysis of glaciolacustrine rhythmites and postglacial marine sediments is another important contribution to a little-known area which requires further geologic study, as a wide spectrum of diverse sedimentary environments are represented here.

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GLACIAL GEOLOGY OF THE
BEBENSEE LAKE MAP-AREA, (NTS 86 M)
GREAT BEAR LAKE REGION NORTHWEST TERRITORIES

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INTRODUCTION

The Bebensee Lake Map-area is immediately north of the northeast arm of Great Bear Lake, bounded by Lat. 67° - 68° N, Long. 118° - 120° W. It is mainly a treeless plain, although small spruce may be found in widely scattered protected sites. The woody plants are mainly 1- to 1.5-m-high dwarf birch and willow. The northernmost spruce trees seen in the east-central Bebensee Lake Map-area were on a seemingly unprotected knoll at 370 m elevation alongside a large 'icing' or aufeis area at Lat. $67^{\circ}21'$ N, Long. $118^{\circ}19'$ W. The treeline lies some 10 km to the south of these coordinates and below 290 m elevation. Here, near a lake with an elevation of 259 m, spruce trees are relatively common with heights of 5 to 7 m. They have a diameter of 10 to 12 cm a metre above ground. But interspersed with these, and on higher tree-free ground, there are old weathered stumps and logs up to 30 or 35 cm in diameter. These observations indicate a modern climatic deterioration. The 'icing' area referred to above, and other smaller ones, were frequented by caribou during the moderately warm August days.

The broad regional pattern of ice-flow features, striae, eskers, and meltwater channels in Great Bear basin presents many problems as regards the sequence and implications of the glacial events. East of Great Bear Lake there is a relatively orderly pattern of ice-flow features, striae, and eskers believed to indicate glacier recession toward the distant, long-recognized "Centre of the Keewatin Glacier" (Tyrell, 1897, p.178-9); the loci of the last active remnant of ice was named the Keewatin Ice Divide (Lee et al., 1957; Lee, 1959). But elsewhere around Great Bear Lake, the ice-flow pattern is complex with southwest, west, northwest and north-trending lineaments. The most puzzling pattern is that of the north-trending flutings, drumlinoid ridges and crag-and-tail features in east-central Bebensee Lake Map-area.

REGIONAL CONSIDERATIONS

The regional ice-flow pattern is shown on the surficial geology reconnaissance map of Craig (1959) and the Glacial Map of Canada (Prest et al., 1968). Craig showed that northward ice flow occurred only east of 119° longitude between McTavish Bay of Great Bear Lake and Bebensee Lake ($67^{\circ}30'$ N, $118^{\circ}30'$ W). A few kilometres north of Bebensee Lake the north-trending features are cross-cut by east-west trending flow features, but examination of airphotos reveals that the northward ice flow did extend at least as far as $67^{\circ}45'$ N before being displaced by west flowing ice (Prest et al., 1968), and Craig observed striations indicative of north flow a short distance north of 68° N latitude. The band of north-flowing ice has a width of 35 to 45 km (Fig. 1). It has a very sharp western boundary, but on the east the flow features merge gradually with northwestward-flowing Keewatin ice. The band of northward flow may well be associated with that shown by Craig (1959) only 15 km farther northeast, which flowed toward Dolphin and Union Strait and thence into Amundsen Gulf. If so,

the advance of Keewatin ice responsible for the northward flow may be an integral part of the Late Wisconsinan advance which reached into, but not beyond, Amundsen Gulf (Vincent, 1982, Fig. 11; Vincent 1984, Fig. 5; Prest, 1984).

The last active ice north of Bebensee Lake appears to have swept westward around the northern end of the receding northward-flowing ice. (It is believed that these two ice-flow trends are intimately related, both stemming from the advance of Keewatin ice from the upland to the east.) This westerly flowing (younger) ice, on encountering a highland in the northwest corner of Bebensee Lake Map-area, was deflected to the southwest and south-southwest. It was, of course, also deflected northwestward around the highland toward Amundsen Gulf. This 'younger' ice was responsible for construction of a nearly east-west morainic ridge in the west-central part of the map area. Shortly before this event, the ice had apparently advanced some 10 to 25 km beyond the morainic ridge as is indicated by the pattern of 'parallel ice-flow features' (Prest, 1968) including directional crag and tail features. It is uncertain whether an 'older' Dismal/Bebensee ice lobe had vacated this area or was merely reoriented by the renewed ice flow from the north. In either case, reactivated Great Bear Basin ice still occupied the low area to the south since high-level glacial pondings and lakes were present in Bebensee Lake Map-area as the younger ice receded.

The terrain north of the morainic ridge is mainly hummocky ground moraine, but in places near the ridge a subdued parallel ice-flow pattern is still discernible; this has the same orientation as that south of the moraine (Fig. 2). Thus it appears that the morainic ridge indicates a minor readvance of the ice while the ice of its own maximum advance was still in place, or else this was actually somewhat older remnant ice. Certainly the steep 30 to 60 meter-high south face of the moraine, with some slump features along it, demands confining ice on its south side at the time of its formation. The north side of the ridge slopes to the north along the greater part of its length. Both large, massive and small, discrete eskers terminate at the moraine. The escaping meltwaters had to be ice-confined at the ridge and flowed off to the west. Thus the morainic ridge, formerly regarded as an end moraine, may represent an ice-confined channel or crevasse within the terminal zone of the ice of the last glacier advance in the region or between that ice and somewhat older ice.

The morainic ridge is mainly composed of sandy materials, but gravel and cobble components were observed, as were large boulders and blocks of local bedrock types in one place at the base of the ridge. Though readily mappable for over 40 km, the moraine displays a number of high knobs or kames with intervening low areas and gaps, and hence might be considered a kame moraine. The ridge has been terraced and beached by glacial lake waters up to 475 or 480 m elevation with only a very few knobs protruding above the lake limit. But there is an



FIGURE 1: Drumlinoid ridges and glacial fluting in southeastern Bebeness Lake Nap-area; note the crag-and-tail features indicative of north-flowing ice; deltaic features in northeast corner of photo are part of an interlobate complex formed in lake at 330 m elevation; the delta at the northern end of the large lake was constructed by the north-flowing ice into a lake at 385 m; 67°16'N, 118°30'W; NAPL A22419-148.

FIGURE 2: 'End moraine' in west-central Bebeness Lake area formed by south-southwest-flowing ice perhaps in a crevasse; note the drumlinoid features south of the ridge and traces of similarly oriented features beneath hummocky ground moraine north of ridge; the moraine displays beaches, terraces and spits up to elevation 480 m; 67°28'N, 119°15'W; NAPL A22412-43.



absence of ice-contact deltas along the ridge.

Eastward the morainic ridge merges -- at $67^{\circ}28'N$, $119^{\circ}08'W$ -- into a markedly different complex of esker, kame, kame terrace, outwash or subwash, and ice-contact deltas (Fig. 2). It no longer represents an ice-confined channel at right angles to the Keewatin ice-flow trend whose outlet is some 50 km to the west, but rather an orifice between older stagnant ice and the margin of eastward-receding ice lobes with meltwaters debouching into glacial pondings and lakes at about 480 m elevation (Fig. 3). Farther east, where the ridge complex crosses the Greenhorn River at $118^{\circ}46'W$, the orifice is clearly between the receding north-flowing and slightly younger west-flowing ice masses. The ridge of glaciofluvial deposits is thus interlobate. The most pronounced delta complex was built where the large esker system, extending westward from Dismal Lakes Map-area, joined the ridge complex at $118^{\circ}20'W$. Here a terrace and pitted delta ($3000 \times 1000 \times 20$ to 30 m) was formed between 410 and 400 m elevation. At the same time, a disjointed system of deltas, at the same elevation, was formed along the northern margin of the north-flowing (southward receding) ice. Clearly, glacial lake waters were present along the northern edge of the southward-receding ice (Fig. 4). Lake drainage was to the west, but the margins of the lake are ill-defined. As the ice-margin orifice receded southeastward, the debouching meltwaters formed ice-contact deltas into ever-lowering lake levels. From the deltas at $67^{\circ}28'N$, $119^{\circ}08'W$, southeastward along the ridge to the boundary between the Bebensee and Dismal Lakes map-areas at $67^{\circ}15'N$, $118^{\circ}30'W$, a distance of about 55 km, the ice-contact deltas and hence the glacial lake levels lower from 480 ± 5 m to 275 ± 5 m. At about the latter level and as the ice receded south to free the Dease River valley, Glacial Lake McConnell (Craig, 1965) entered the southeast corner of the map area and extended into the southern part of the Dismal Lakes Map-area at about 285 m. In the latter area strandline features at 290 m over a broad area are related to Glacial Lake McConnell (St-Onge and Dredge, 1985). The generalized glacial geology of the Bebensee Lake Map-area, as determined from air photo studies and limited ground observations in the east-central part of the areas, is shown in Fig. 5.

The successive positions of the confining western ice plugging the Great Bear Lake Basin and the lake levels related to them are uncertain. Some outwash-alluvial systems constructed deltas into standing bodies of water, and a few remnant ice-contact type deltas are discernible on airphotos. Temporary lake stands in southwestern Bebensee Lake Map-area were at elevations of about 480-460 m west of Greenhorn River, and then at 430-420, 370-360 and 340-330 m, lowering southeastward and also south toward Great Bear Lake. Finally, as the ice receded from Great Bear Lake itself, it was replaced by Glacial Lake McConnell. Its approximate strandline elevations, inland from the north shore of Great Bear Lake, increased from 206 m at longitude $122^{\circ}30'W$ (Craig, 1965) to 229 m at $120^{\circ}15'W$ (Fulton, pers. comm.) to 244 m at $119^{\circ}30'W$ and 285 m at $118^{\circ}15'W$; it occupied the southwest corner of Dismal Lake Map-area at 290 m (St-Onge and Dredge, 1985). The increase in elevation of the Lake McConnell strandline toward the east-northeast is due to isostatic adjustment of the land surface following deglaciation, but the precise direction of maximum tilting is not known.

GENERAL GEOLOGY

Bedrock

The western half of a reconnaissance map of the north-central District of Mackenzie (Map 18-1960) included the Bebensee Lake Map-area. According to the descriptive notes (Fraser, 1960) the rocks in the Bebensee area include sandstone, quartzite and conglomerate, grey to red dolomite with stromatolitic strata, and minor shaly dolomite of the Hornby Bay Group; these are overlain by dark brown to purplish basaltic flows - commonly amygdaloidal - comprising the lowest part of the Coppermine River series; the sediments and lavas are intruded by numerous diabase dykes. All of these rocks are Proterozoic in age.

Immediately east of the Bebensee Lake Map-area is the Dismal Lakes Map-area, for which a report and coloured geological map have been published (Baragar and Donaldson, 1973). Here there are both older and younger sediments and felsitic and granitoid intrusive bodies; the latter are readily recognizable as erratics in east-central Bebensee Lake Map-area.

The complexity of bedrock types and structures south of Dismal Lakes Map-area and east of Great Bear Lake is shown on a recently published coloured geological map (Hoffman, 1984).

Glacial Geology

Apart from a brief visit to the morainic ridge in west central Bebensee Lake Map-area, which included a flight south of the ridge and examination of a crag and tail feature there, recent field observations were limited to the area of south to north 'parallel ice-flow features' (Prest, 1968) and the ice-contact ridge complex in the general vicinity of Bebensee Lake. The south to north ice-flow features constitute a remarkably striking linear pattern (Fig. 1). From a cursory study of the topographic and reconnaissance surficial geology maps, the pattern would appear to represent an ice flowage that was independent of any other in the region. The flow pattern has a very sharply defined west boundary against features indicative of southwestward ice flow. But eastward and southeastward the south to north ice-flow trends merge gradually with features left by northwestward-flowing Keewatin ice. The flow patterns strongly suggest that the north and northwest-flowing ice masses were intimately related, and that the latter gradually predominated as the former receded southward; the main Keewatin flow thus swung westward in the Bebensee Map-area.

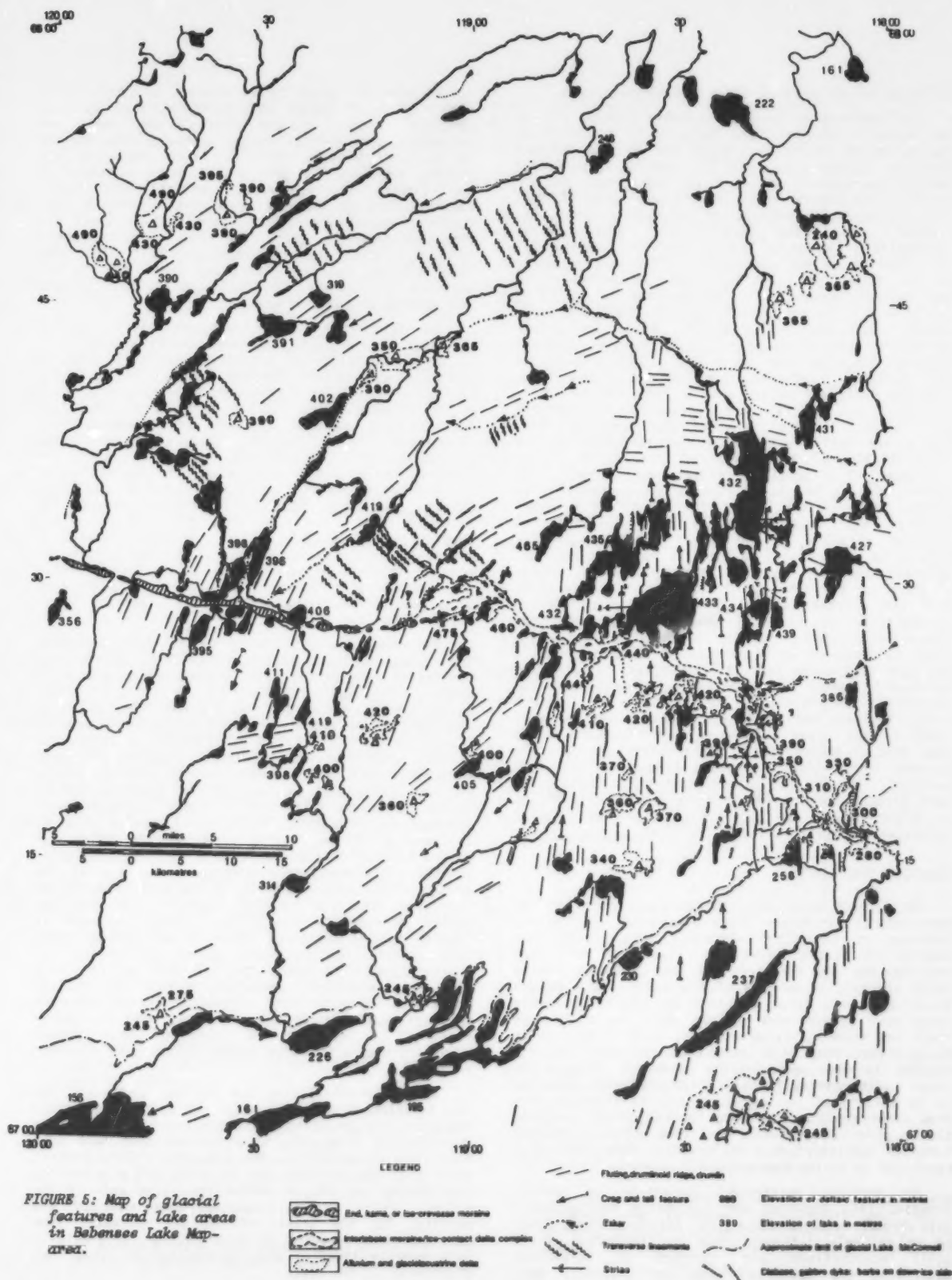
In essence, the generally westward-trending Keewatin ice flowing off the Keewatin upland met resistance in the Great Bear Lake basin and was forced to turn sharply northward in the vicinity of $119^{\circ}W$ longitude. It was obviously confined on the west by a massive glacier and flowed north across terrain ranging from 155 to 550 m in present-day elevation. Presumably the glacier mass in the low-lying Great Bear Basin was somewhat reactivated at the same time as the late advance of Keewatin ice from the higher terrain to the east. The northward ice flow served to cut off the southwest flow of a former Dismal/Bebensee ice lobe across southwest Bebensee Lake Map-area; this ice had formerly served as one of the supply routes into the Great Bear Lake Basin. It is probable that the Great Bear ice regimen was drastically affected by the change in direction of the main Keewatin flow, and the life of



FIGURE 3: Ice-contact complex of esker, kame terrace, kame, outwash/subwash, and deltaic features; the east end of the ridge in Fig. 2 joins that in the southwest corner of this photo. Note how the configuration of the outwash/delta mounds denotes the northeastward retreat of the ice margin; 67° 26.5' N, 113° 06' W; MAPL A33812-95.

FIGURE 4: Ice-contact complex south of Rebenasee Lake formed between west-flowing and north-flowing glacier lobes, and deltaic features built somewhat later along the northern edge of the north-flowing ice; deltas at northeast corner of photo were built into a lake at elevation 450 m, those in southern part of photo at 420-410 m; two small eskers formed by northward-escaping meltwater; bedrock strata are at right angles to the ice flow here; 67° 25' N, 118° 40' W; MAPL A33812-38.





at least its northern part was thereby shortened.

The presence of younger westward-flowing ice in northeastern Bebensee Lake Map-area is indicated by the course of two esker systems and by some subdued ice-flow lineations, albeit closely paralleling bedrock structures in part of the area. But in southeastern Bebensee area no east-west parallel ice-flow features are evident on airphotos. There is, however, one major west-draining esker system indicative of a former ice cover. Close examination of airphotos of the ice-contact ridge complex which extends from beyond Bebensee Lake southeastward to the eastern boundary of the map-area (at $67^{\circ}15'W$, $118^{\circ}00'W$) indicates recession of contemporaneous southern and eastern ice masses. Also, ice-contact deltas on both sides of the ridge complex commonly display slump features indicative of the removal of the formerly confining ice. Ground observations provided foolproof evidence of a late ice cover; in several places glacial striae revealed that westward ice flow had taken place after the northward flow. Striae are well preserved on diabase dykes. At one site ($67^{\circ}22.8'N$, $118^{\circ}19.9'W$) at the south end of a segment of a prominent diabase dyke trending at north $10^{\circ}E$, the north-flowing ice scoured the eastern side of the dyke. The 10 to 15 m near-vertical wall was remarkably well polished and striated parallel to the dyke side; younger ice flowing at right angles to the dyke had scratched the polished surface, leaving short disjointed striae directly up the dyke wall. Also striations trending 270° and related to the younger ice were observed on quartzite bedrock on the north side of the esker system about 4000 m to the northeast of this dyke site. Evidence of westward flow is also evident in the form of the bouldery, blocky western sides of the north-south dykes as compared to the sharper and generally near-vertical but rounded-off eastern sides; boulders and blocks of diabase along the west sides of the dykes in places form a sloping apron some tens of meters broad. Striae sites are commonly present where joints provided breaks in the dyke profiles. Though both north- and west-flowing ice went through these breaks, it is possible in places to differentiate the two events; the west striae cross-cut the north striae.

SURFICIAL MATERIALS

As already mentioned, the morainic ridge in the west-central Bebensee Lake Map-area is mainly composed of sand-sized materials, but gravels and cobbles are evident where the ridge knobs have been both beached and terraced, and some gravelly, hooked spits are evident on some of the knobs. Only in one place were very large boulders and blocks observed; these were at the base of the east end of a high ridge and formed a pyramid on the basal slope. It is assumed that they represented the core of the ridge exposed by drainage escaping southward through the break in the ridge. The sandy deposits were observed to be stratified in only one adequate exposure, but from a lengthy traverse along the ridge, and airphoto studies, it is believed that the ridge is mainly composed of stratified materials. This was also concluded by Fulton (pers. comm.).

As earlier mentioned, beginning in central Bebensee Lake Map-area and continuing southeastward into Dismal Lakes Map-area, the morainic ridge takes on an entirely different form and character. While still ice-confined and composed of stratified materials, the meltwater flow was assuredly more

torrential. The ridge system appears to consist of a number of kames which have been terraced by ice-confined waters flowing toward the west and northwest; the terraces merged westward into subwash or outwash deposits that are both channeled and pitted, and these in turn either merge westward or are flanked by ice-contact deltas. The latter commonly display slumping along their formerly ice-confined sides. Within the ridge complex there are also some discrete eskers. The kame - kame terrace - outwash - delta complex repeats itself along the ridge system, each unit becoming lower towards the east and southeast as the confining ice both thinned and receded in those directions.

The kame or kame-like knobs are mainly composed of large angular and subrounded blocks and boulders of white quartzite and sandstone, and dark greenish diabase, with rather limited foreign boulders and cobbles. The largest blocks and boulders are up to 2 m in diameter, but average about 1 m. The kame terraces and subwash deposits are somewhat less blocky and more bouldery. The esker ridges within the complex are distinctly more bouldery and cobbly, and foreign stones are more common. The deposition of the kame, kame-terrace and esker parts of the ridge complex and their declining elevations towards the east and south clearly indicate that the ridge was constructed in sections as the ice masses receded to the south and east. Nevertheless, erratics within the complex indicate transport from south-central Dismal Lakes Map-area, judging by the bedrock map of that area (Baragar and Donaldson, 1973) - perhaps 50 to 75 km distant.

The esker system entering the Bebensee area from the Dismal Lakes area at $67^{\circ}27'N$, $118^{\circ}W$ and continuing west to the main morainic ridge system at about $67^{\circ}24'N$, $118^{\circ}20'W$ is a more typical esker. It forms a discrete narrow sinuous ridge composed of moderately rounded cobble and gravel-sized stones with but a scattering of quartzite blocks. Granitoid and porphyritic stones were more noticeable here than anywhere in east-central Bebensee area. They were no doubt derived from the bedrock masses in nearby west-central Dismal Lakes Map-area (Baragar and Donaldson, 1973). The low esker ridge is bordered on its north side by a 500- to 800-m-wide washed zone which appears to be a shallow, subglacial channel that immediately superseded the use of the more confined esker channel. On the south side of the esker, as the ice receded eastward, the escaping waters emerged to deposit a braided outwash or subwash system expanding westward and southwestward into deltaic bodies; these appear to outline the lobate form of eastward-receding ice pulling back from the interlobate morainic ridge. The glacial meltwaters debouched into ephemeral glacial lakes or pondings held up on the south by the southward-receding ice and on the north by high ground bordering the south side of the interlobate morainic ridge.

As the north-flowing ice receded southward from the interlobate morainic ridge in southern Bebensee Map-area, the escaping meltwaters were first carried northward by three tiny esker systems that have been observed on airphotos. One debouched into a glacial ponding at the moraine edge at about 440 m elevation and another at 420 m, while the third entered the 420 m ponding or lake some distance west of the moraine. But the great bulk of the meltwater from this ice lobe appears to have escaped subglacially into rapidly lowering glacial pondings or lakes that discharged to the west.

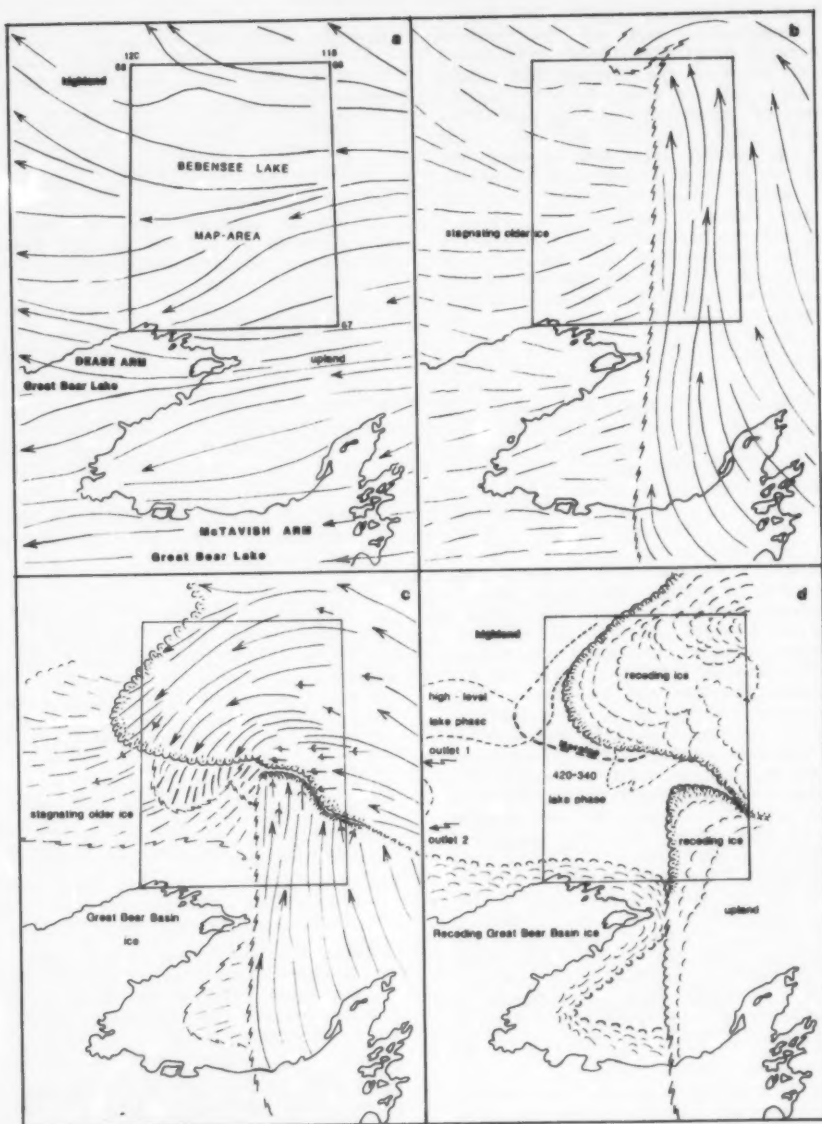


FIGURE 6a-d: Sketch maps showing problematical phases of glacial activity with ice margins, and glacial lakes:

- i) Arrows and long lines = generalized ice-flow trends; ii) Scalloped lines = ice margins, known and inferred; iii) Striae symbol = recorded striae and other ice-flow indicators; iv) Zig-zag lines = ice confluence zones; v) Short dashed lines = shore lines of glacial lakes - mainly inferred.

a) Location maps, and sketch of an early and main glacier movement (Kewatin ice).

(b) North-trending ice flow cross-cutting the earlier movement and thereby cutting off its supply sources and promoting its stagnation.

(c) North-flowing glacier receding and being replaced by west- and southwest-flowing glacier with construction of an 'end moraine' in west-central Babensee Lake Map-area.

(d) Interlobate kame-outwash-delta complex formed between the receding west-flowing and north-flowing glacier masses, and showing glacial lake phases held-up by the receding Great Bear glacier.

The areas of obviously washed bedrock and drift, as seen on airphotos and checked against the modern 1:50,000 scale topographic maps of southeastern Bebensee area, revealed great elevational differences that are not in accord with the falling lake levels that also washed parts of the area. In other words, there is evidence of extensive washing and even removal of the drift cover that is well above the local altitude of the several lakes that bordered the southward-receding ice margin. A bouldery to blocky lag deposit of mainly quartzite, sandstone and diabase blocks and boulders mantles the bedrock wherever extensive washing has taken place. The drift mantle everywhere is probably thin - in the order of 1 to 2 m. The unwashed till of the ground moraine areas was observed to be sandy, but no analyses are available.

ECONOMIC CONSIDERATIONS

Neither the Proterozoic sedimentary strata nor the numerous diabase dykes in the Bebensee Lake Map-area appear to be of economic interest. There is, however, a band of amygdaloidal basalt within the Bebensee area that elsewhere is known to contain minor amounts of native copper as well as chalcocite, covellite and bornite (Fraser, 1960). Older Proterozoic rocks, including intrusions, occur to the east and southeast of the Bebensee Lake Map-area, and these do have economic potential. The known mineral occurrences and references to older work are included in Fraser's Descriptive Notes on Map 18-1960. The importance of copper deposits in Dismal Lakes Map-area is evident from the descriptions given on the Dismal Lakes Map-area (Baragar and Donaldson, 1973) and several references therein. It is in connection with the occurrences to the east and southeast of Bebensee Lake Map-area that possible glacial dispersion may be important; boulder or mineral-tracing work within or around the Bebensee Map-area must take account of the successive and diverse trends of glacier flow (Fig. 6). As mentioned earlier, glacial erratics are present in east-central Bebensee Lake area where they were mainly dispersed by the last active and westward-flowing ice, but the input of the north-flowing ice is presently uncertain. Foreign stones are mainly contained in the esker entering the area from the Dismal Lakes Map-area and ending at the moraine-kame-delta complex, and within parts of the ridge; they are uncommon in the ground moraine which is largely composed of local quartzite and sandstone, with a sprinkling of diabase.

ACKNOWLEDGEMENTS

Dr. D.A. St-Onge of the Geological Survey of Canada kindly proffered the use of his camp at Dismal Lakes as a base to examine the surficial features in the east-central Bebensee Lake Map-area. The field work was greatly expedited by Dr. St-Onge and his party members. Dr. St-Onge was also instrumental in preliminary arrangements as regards to travel funding through Dr. W.A. Padgham of the Department of Indian and Northern Affairs. The writer is extremely grateful to Drs. St-Onge and Padgham for their making the field study possible. He also wishes to acknowledge several useful discussions with Dr. R.J. Fulton of the Geological Survey of Canada, who visited the region in 1968 as a member of the Survey's Operation Norman and prepared air-photo interpretation maps of parts of the area.

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GEOLOGY OF
NORTHWESTERN BRODEUR PENINSULA,
BAFFIN ISLAND, N.W.T.

by

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ABSTRACT

On northwestern Brodeur Peninsula, Baffin Island, the Brodeur Group consists of the Baillarge and Cape Crauford formations, a succession of Ordovician and Silurian carbonates deposited in shallow, low-latitude tropical seas. The Baillarge Formation is 514 m thick and comprises members A and B. Member A (140 m) consists of dolostone interbedded with limestone (in part breccia) while member B (374 m) consists of thick massive dolomitic limestone containing minor solution breccias. The overlying Cape Crauford Formation, which is up to 360 m thick, is divided into lower and upper members. The lower member is 130 to 208 m thick and consists of limestone (in part solution breccia) interbedded with variously dolomitic limestone and dolostone. The upper member is 152 m thick and is divided into units "a" and "b". Unit "a" (110 m) consists of massive buff dolostone that has been mapped over most of the peninsula. Unit "b" (42 m), which overlies unit "a", is lithologically similar to the lower member of the Cape Crauford Formation. An unconformity between the upper and lower members of the Cape Crauford Formation is recognized throughout the area and represents a period of subaerial exposure in the Llandoveryan. This exposure generated the paleokarst breccias of the lower member of the Cape Crauford Formation. It is suggested that the breccias were formed by the dissolution of limestone rather than evaporites, an interpretation that gives the area a high lead-zinc mineral potential.

INTRODUCTION

Brodeur Peninsula, north of 73°0', is a plateau, most of which lies between 150 and 300 m above sea-level, although some areas in the eastern part of the peninsula range above 450 m (Fig. 1). Maximum elevations of 580 m occur near Sargent Point and near the east coast. In a general way the regional northwest dip of the Brodeur Group strata is reflected by a northwest-sloping land surface (Figs. 1 and 6). Numerous streams and rivers transect the surface to form a rectilinear drainage pattern. Most of the plateau is covered by surficial deposits of Pleistocene and Recent age, aspects of which were considered by Craig (1965) for southern Brodeur Peninsula. Consequently, the best bedrock exposures occur in coastal cliff exposures although some bedrock crops out along river valleys.

PALEOGEOGRAPHIC SETTING

Northern Brodeur Peninsula was situated on a shallow carbonate shelf about 200 km southeast of the Franklinian Geosyncline (Nentwich and Jones, 1984). During deposition of the Brodeur Group between Late Ordovician and Silurian times, the study area drifted south as part of the North American Plate from 18° to 12° north of the equator (Irving, 1979). A paleomagnetic study of the Pridolian Peel Sound and Lochkovian Snowblind Bay formations on nearby Somerset, Prince of Wales and Cornwallis islands by Lapointe and Dankers (1981) showed that these sediments were magnetized 10° north of the equator.

PREVIOUS GEOLOGICAL INVESTIGATIONS

The geology of Brodeur Peninsula has received relatively little attention, even though it has been visited by many people. Port Bowen was visited and named by Parry in 1819 who noted that "The cliffs on the south side of this bay, to which I gave the name of Port Bowen, resemble, in many places, ruined towers and battlements; and fragments of the rocks

were constantly falling from above: (Parry, 1835, p. 52). It seems that Parry was referring to the breccia bodies which are so apparent in this area. Fossils collected by Parry during his third expedition were described by Jameson (in Parry, 1826) and later by Lee (1912). Ross (1835) noted that the strata at Cape York consist of limestone and Haughton identified a colonial coral specimen collected at Cape York (in McClintock, 1859). In 1911, Tremblay and Lavole explored parts of the peninsula (in Bernier, 1912), and corals collected in 1924 at Stanley Point during the Fifth Thule Expedition were described by Teichert (1937).

Comprehensive stratigraphic investigations began when Blackadar (1956), Lemon (1956), and Lemon and Blackadar (1963) named and described the type section of the Baillarge Formation at Baillarge Bay opposite Ship Point (Fig. 1). The strata between Port Bowen and Sargent Point were later investigated by Norris (1963) who delineated three informal stratigraphic units, namely (1) a lower limestone, (2) a middle brecciated limestone, and (3) an upper limestone. Trettin (1965a,b) measured sections at Cape Crauford, Sargent Point, and Jackson Inlet and revised the stratigraphic framework established by Norris (1963) because (1) part of the lower limestone unit of Norris (1963) is part of the Baillarge Formation and (2) the strata overlying the Baillarge Formation include solution breccias and are therefore lithologically distinct from the unbrecciated strata of the Read Bay Formation on Cornwallis Island.

Additional information on the geology of the peninsula is contained in industry reports released from confidential status (reports prior to 1977). These include Heise (1970), Howell (1970), Mayr and Harding (1971) and Miall and Harding (1971). Mineral exploration on Brodeur Peninsula has been conducted by Cominco Ltd. (W. Gibbins, DIAND, pers. comm., 1982) and Monopros Canada Ltd. (D. Boucher, Hudson Bay Mining and Smelting Co. Ltd., pers. comm., 1982). Of peripheral interest is geophysical work, especially marine seismic reports. These may prove to be of importance to regional stratigraphic correlation and include Anonymous (1972), Ball (1973), Ball

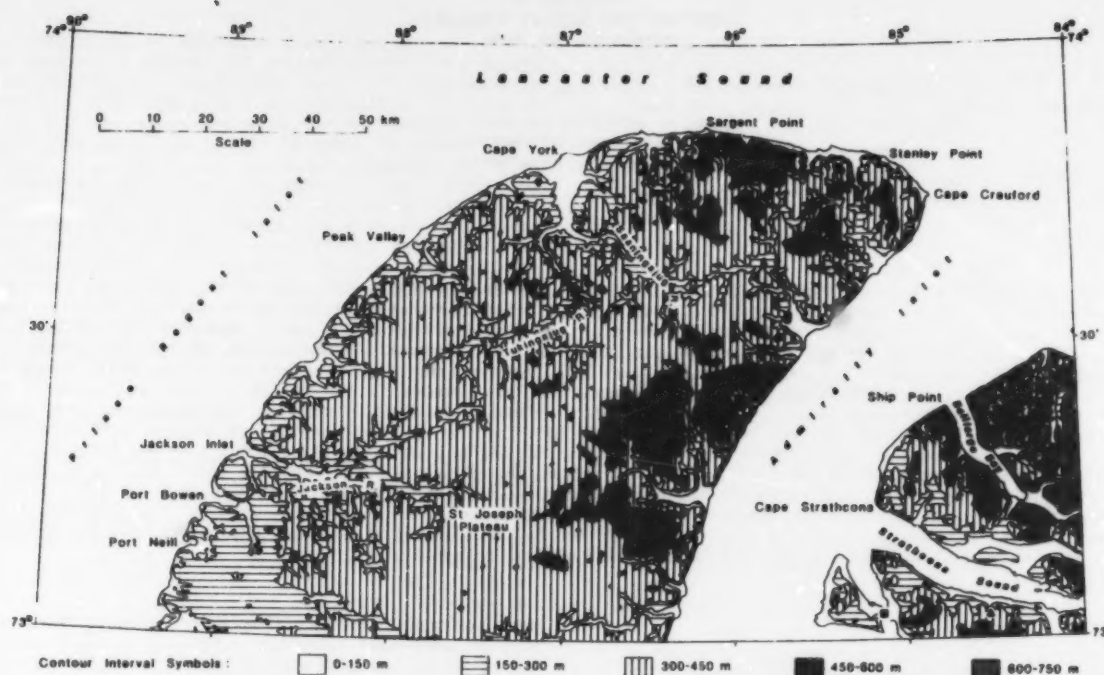


FIGURE 1: Topography and main geographic features of northern Brodeur Peninsula.

et al. (1973), Berkhout (1973), Lack (1971, 1972a,b,c), Pflueger (1974) and McGrath and Fraser (1981). An aeromagnetic survey was conducted by the Department of Energy, Mines, and Resources during the summer of 1984 (Dr. P. Hood, EMR, per. comm., 1984). Previously published reports by the authors include Nentwich and Jones (1983, 1984a,b).

STRATIGRAPHY

The succession exposed on Brodeur Peninsula comprises in ascending order the Ship Point, Baillarge and Cape Crauford formations (Fig. 3) The latter two constitute the Brodeur Group as named by Trettin (1965b).

Ship Point Formation

The uppermost 38 m of the Ship Point Formation occur in sections M and MX (Figs. 2, 6, 7) southwest of Cape Crauford. There it is a homogeneous, mottled, medium crystalline dolostone that contains chain corals, gastropods, nautiloids, echinoderm ossicles and conodonts.

Baillarge Formation

Definition

The formation was named by Blackadar (1956) and described by Lemon (1956) and Lemon and Blackadar (1963) who designated a poorly exposed 119.5-m-thick section at Baillarge Bay as the type section. The formation was extended upward by Trettin (1965a, p. 160) "...to the first significant and mappable change

in lithology." Trettin (1965b) designated a reference section about 25 km southwest of Cape Crauford where he divided the formation into informal A and B members, a division followed in this study.

Member A

For the present study member A was measured at sections MX, N, and P (Figs. 2, 5).

Lithology

Since all gradations between pure limestone and pure dolostone occur, the following terminology is used: limestone (0-10% dolomite), dolomitic limestone (10-50% dolomite), calcareous dolostone (50-90% dolomite), and dolostone (90-100% dolomite). The basal 60 m are covered and overlain primarily by dolostone with some interstratified limestone and rare thin beds of argillaceous and shaly limestone. There is only one stratified breccia in member A (Figs. 3, 7), but it is of interest because of its similarity to those in the overlying Cape Crauford Formation. The breccia occurs in pure limestone at section MX (Figs. 2, 5) and overlies stromatolitic structures that are about 1 m wide and 20 cm high. The breccia is inversely graded with respect to clast size and stylolites are common throughout. At section N, situated 350 m to the northeast of section MX, stylolitic limestone occurs rather than breccia. Sixty metres of mainly fine to medium crystalline dolostone overlies the breccia. Although fossils are rare and poorly preserved in the dolostone, gastropods, disarticulated brachiopods, stromatolites, and echinoderm ossicles do occur. In addition, one trilobite pygidium was found in the scree and conodonts were recovered.

Basal Contact

The contact with the underlying Ship Point Formation is covered at the reference section. Trettin (1965b, p. 11) suggested that this boundary is a regional disconformity because, at various localities on northern Borden Peninsula, the Baillarge Formation is anomalously thin and there is a solution zone at the top of the Ship Point Formation. There is also an abrupt change in lithology at the contact between the Ship Point Formation and the recessive strata of member A.

Distribution and Thickness

Member A is 140 m thick and was measured in sections MX, N, and P (Figs. 2, 5).

Age

No fossils were collected from member A by Trettin (1965b). However, the member occurs above early Middle Ordovician rocks and below the late Middle and/or early Late Ordovician rocks of member B of the Baillarge Formation. For this reason a late Middle Ordovician age was assigned by Sinclair (in Trettin, 1969). This is in keeping with a late Llanvirnian to early Caradocian age determined by Barnes (1973) for conodonts derived from the upper part of the Ship Point Formation.

Member B

Lithology

Member B is more fossiliferous than member A and consists of massive to thick-bedded mottled dolomitic

limestone and very minor interstratified pure limestone. A local breccia occurs 187 m above the base of the member, 30 m below a finely crystalline dolostone containing ripple marks and dessication cracks (Fig. 7). A stratified solution breccia in a thick-bedded mottled limestone occurs 293 to 310 m above the base of the member. The clasts in this breccia are angular, have a mode between 5 and 10 cm, and range up to 15 cm in size. The clasts are inversely graded with respect to size; thus, 4 to 5 m above the base of the breccia the clasts have a modal size of about 20 cm. Extensive high-amplitude stylolites are very common. The breccia is laterally continuous over at least 100 m at section L, but it was not found in the cliff 1 km to the southwest. Fossils are abundant, diverse, and include orthoconic nautiloids, rugose, chain, and other colonial corals, brachiopods, echinoderm ossicles and stems, stromatoporoids, stromatolites, ostracods, and conodonts.

Basal Contact

The basal contact with member A is taken at the base of a massive dolomitic limestone succession.

Distribution and thickness

Member B is mainly confined to the northeastern part of the map area between Sargent Point and the reference locality, 25 km southwest of Cape Crauford (Figs. 5, 6). The member is 356 m thick at section L where almost all of the member is exposed. Elsewhere it is less well exposed because of the regional northwest dip of the strata (Figs. 5, 6). Thus, the member is either partly below sea-level or is mostly covered by talus at the base of the cliffs. The upper

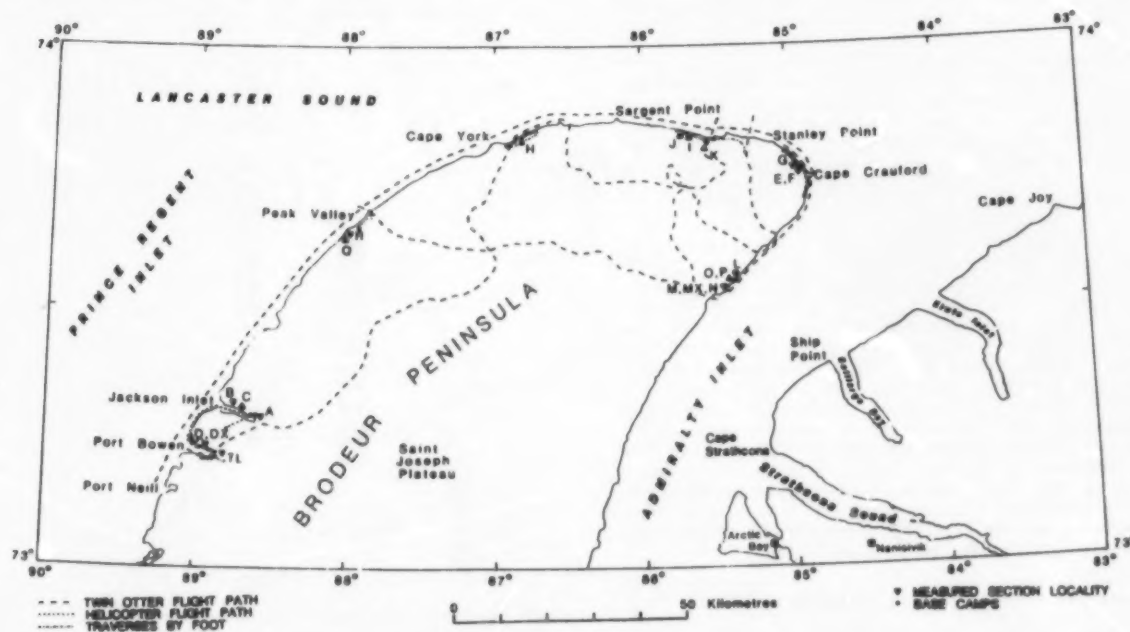


FIGURE 2: Sites of base camps, measured sections, and flight paths along which aerial photographs were taken.

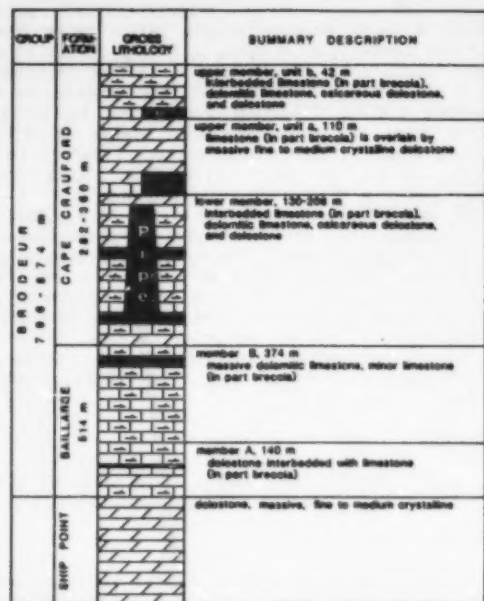


FIGURE 3: Composite lithostratigraphic column of measured strata. Breccia bodies are shown in solid black.

half of member B is well exposed in the northern coastal cliffs between Sargent Point and Stanley Point. It is less well exposed in the interior of the peninsula, although it does crop out in cliff exposures along some of the major streams, especially the Tukingajug and Saaningajug rivers (Fig. 1). The upper 30 m of the member is also exposed at Port Bowen and Jackson Inlet. Prominent units up to 50 m thick in the upper part of the member are traceable without noticeable thickness change in cliff photographs for up to 40 km between Sargent Point and Stanley Point.

Age

The basal 87 m of member B is late Middle to early Late Ordovician in age (in Trettin, 1969). From 195 to 295 m above the base there are numerous brachiopods including *Rhipidium* (Bolton, in Trettin, 1969). These were subsequently reassigned to the middle Llandoveryan genus *Virgiana* by Berry and Bourcot (1970). Conodonts from a 33-m-thick interval collected somewhere near the base of an interval 104 to 193 m above the base of the member are Richmondian or possibly Late Maysvillian in age (Barnes, 1973). Silurian fossils occur from 152 to 346 m above the base (Bolton, in Trettin, 1969). A study of conodonts processed for the present study suggests that the Ordovician-Silurian boundary occurs in the upper part of member B of the Baillarge Formation (B. Chatterton, University of Alberta, pers. comm., 1985).

Cape Crauford Formation

Definition

The Cape Crauford Formation is exposed throughout the study area (Fig. 6) and was defined by Trettin (1965b, p. 18) to include "...all lower Paleo-

zoic strata of northwestern Baffin Island overlying the Baillarge Formation." The type section was defined between 2.4 km south and 3.2 km northwest of Cape Crauford where the informal members A, B and C were delineated (Trettin, 1965b). Members A and B are lithologically similar and are thus difficult to distinguish outside the type locality. Consequently, they are herein grouped together and termed the lower member. Member C of Trettin (1965b) is herein termed the upper member.

Lower Member

Lithology

At the type locality the lower member consists of medium brownish-grey limestone or dolomitic limestone interbedded or intergraded with buff dolostone or calcareous dolostone. From a distance this gives the unit a distinctive striped appearance (Figs. 8c, 9c, 9d).

About 20% of the lower member consists of breccia, developed almost exclusively in limestone. Another 10% consists of stylolitic limestone (Fig. 4) grading to crackle breccia (classification of Morrow, 1982). Unfortunately there are no distinctive marker beds in the lower member that can be correlated throughout the area and lithofacies changes are dramatic even between sections a few kilometres apart.

Fossil content and diversity varies both laterally and vertically in the lower member. At the type section, fossils in the lower 145 m of the member are common, but diversity decreases upwards (Fig. 4).

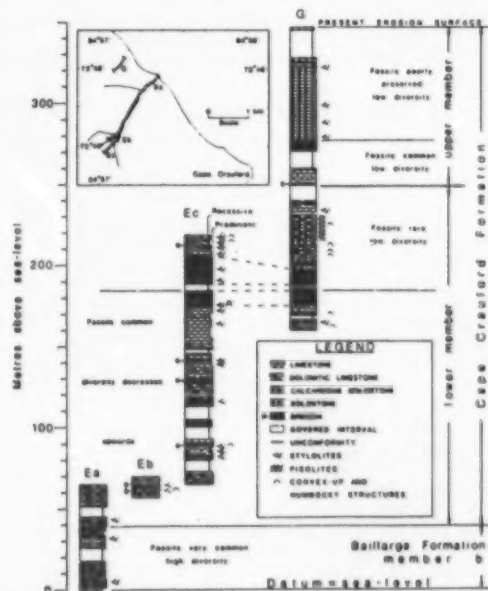


FIGURE 4: Simplified type section of the Cape Crauford Formation. Inset shows where sections were measured.

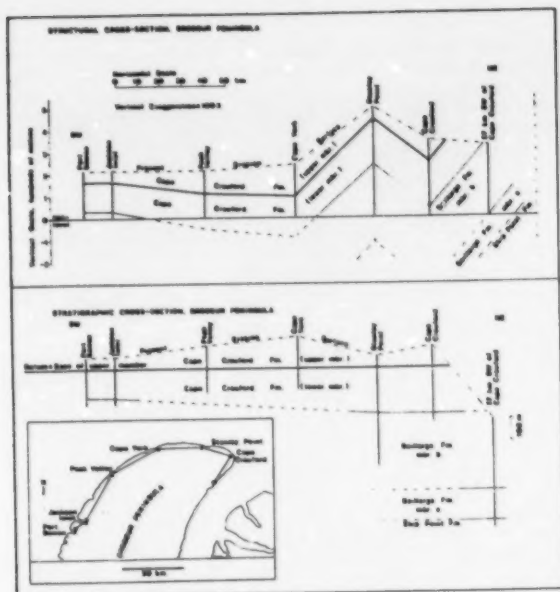


FIGURE 5: Structural and stratigraphic sections, Brodeur Peninsula.

Fossils in the remainder of the lower member, from 145 to 202 m above the base, are rare and have a low diversity. At Stanley Point fossils are less common than at Cape Crauford, but a similar upward decrease in diversity is recognized. The upper 39 m of the member, which is almost barren of macrofossils, is dominated by convex-up and hummocky structures that

resemble stromatolites. The upper 35 m of the lower member at Cape Crauford (Fig. 4) is a complex interbedded lithology of breccia, stylolitic limestone, limestone containing hummocky and convex-up structures, and variously dolomitic pisolitic limestone (Fig. 7). The thickest pisolitic zone, which is 9 m thick, is inversely graded with respect to the size of the pisolites. The upper part of this pisolitic interval is entirely dolomitic. Many of the pisolites are multi-stage types with compound nuclei. Asymmetrical banding is also common and some pisolites are 10 or more times as long as they are wide. Isolated breccia pipes and spectacular collapse structures that may be paleo-dolines extend up to or just below this unconformity (Fig. 8d,8e). At Cape York only the upper 90 m of the member is exposed and except for a few ostracods and brachiopods, the entire section is unfossiliferous. Similar stromatolitic structures are common in the uppermost 33 m. At Peak Valley fossils are rare and much of the section is covered. At Port Bowen and Jackson Inlet the lower half of the member contains diverse fossils, but these are less abundant than elsewhere. Strata in the upper 60 m contain a low population and diversity of fauna.

Basal Contact

The contact with the underlying Baillarge Formation is placed at the base of the first interstratified calcareous dolostone and microcrystalline limestone, a unit generally less than 6.1 m below a sequence of solution breccias (Trettin, 1965b). At Stanley Point the contact is abrupt and conformable. At Jackson Inlet the Baillarge Formation is massive dolostone or interbedded limestone and dolostone, but it can be distinguished from the Cape Crauford Formation on the basis of its massive, mottled character.

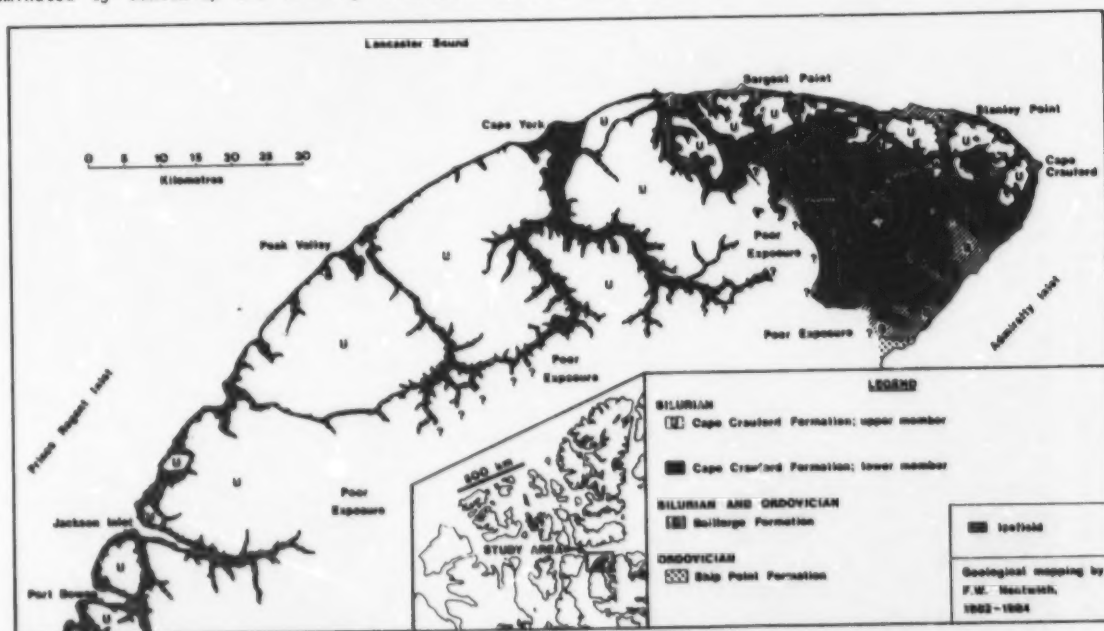


FIGURE 6: Bedrock geological map of northwestern Brodeur Peninsula. Inset shows geographic setting of the study area.

Distribution and Thickness

The lower member is present in all of the measured sections except L (Figs. 5, 6). In addition to coastal exposures it crops out in river valley gorges in the interior of the peninsula. The member is 202 m thick at Cape Crauford, 208 m thick at Stanley Point, but only 135 m thick at Port Bowen. The base of the member dips below sea-level at Peak Valley and at Cape York so that it is not possible to determine the manner in which the member thins between Stanley Point and Jackson Inlet.

Age

Fossils collected from the lower member were identified by Bolton (in Trettin, 1969) who considered them to be Niagaran. Conodonts processed for the present study were considered to be of Llandoveryan age (B. Chatterton, University of Alberta, pers. comm., 1985).

Upper Member, Unit "a"

Lithology

Unit "a" is a massive buff to orange-weathering, coarse-grained, vuggy or mottled dolostone with minor breccia and laminated dolostone (Fig. 3). The unit is further characterised by a thick, dark brownish-grey limestone bed at its base (Figs. 8a, 8b, 8c, 9a). Minor amounts of breccia and laminated dolostone also occur (Fig. 7). At Cape York the unit is 100 m thick (Fig. 5).

With the exception of conodonts, fossils in the unit are generally poorly preserved except in the basal thick limestone bed. At Stanley Point fossils include brachiopods, echinoderm ossicles and solitary corals. Fossils at Peak Valley are diverse and

include corals, brachiopods, echinoderm ossicles and stromatoporoids. At Cape Crauford ostracods, echinoderm ossicles, brachiopods, and orthocones were found and at Cape York the fossil assemblage includes brachiopods, echinoderm ossicles, orthocones, stromatoporoids and corals.

Basal Contact

The upper-member beds, which extend across the collapsed zones of the lower member, are neither downwarped nor downfaulted into them (Fig. 8e). In one case observed in section DX at Port Bowen (Fig. 2), upper-member beds overlie erosionally truncated collapsed strata of the lower member. The basal contact of the unit is unconformable in the Port Bowen area because beds of the upper member overlie collapse structures up to 100 m wide.

Distribution and Thickness

Unit "a" forms the top of the plateau in the western and northern part of the peninsula where it is eroded to various levels. It is distinguished by its orange-weathering colour even in cases where it is poorly exposed in the interior of the peninsula (Fig. 6). This allows accurate mapping where good exposures are few.

Age

Fossils are poorly preserved because of extensive dolomitization. A *Porpites* was collected, however, from 3 to 6 m above the base of Trettin's member C at Cape Crauford (Trettin, 1969). This genus is late Late Llandoveryan to Early Wenlockian in age (Berry and Boucot, 1970). Conodonts recovered during the present study are of Llandoveryan age (B. Chatterton, University of Alberta, pers. comm., 1985).

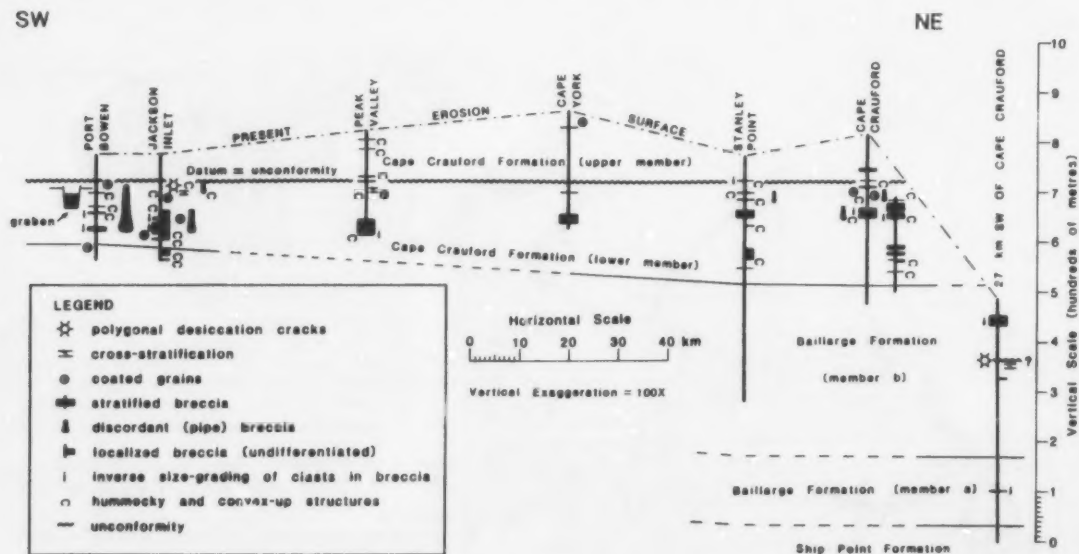


FIGURE 7: Stratigraphic cross-section showing selected geological features.



FIGURE 8a: The upper member (U) of the Cape Crauford Formation. Note the thick limestone bed at its base. Photograph of view to the northwest, taken 5 km southeast of Stanley Point.



FIGURE 8b: The upper member and part of the lower member of the Cape Crauford Formation at section H near Cape York. Note the thick limestone bed at the base of the upper member.



FIGURE 8c: The lower member of the Cape Crauford Formation at the type locality (section F). Field assistant is standing on the top of a buttress 123 m above the base of the lower member.



FIGURE 8d: Small collapse structure just below the unconformity between the lower and upper members of the Cape Crauford Formation, north shore of Port Bowen.

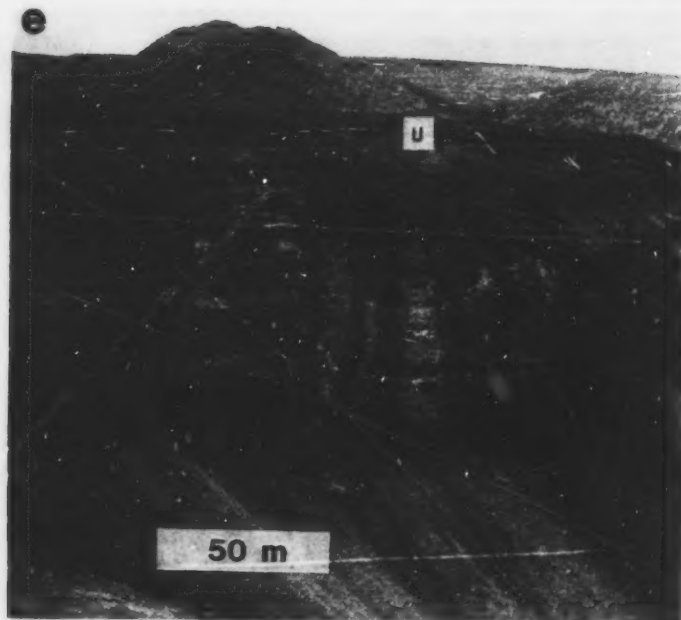


FIGURE 8e: A 110-m-wide breccia body on the south shore of Port Bowen (outlined). The upper member extends undisturbed over these collapse structures and is therefore an unconformity (wavy line).

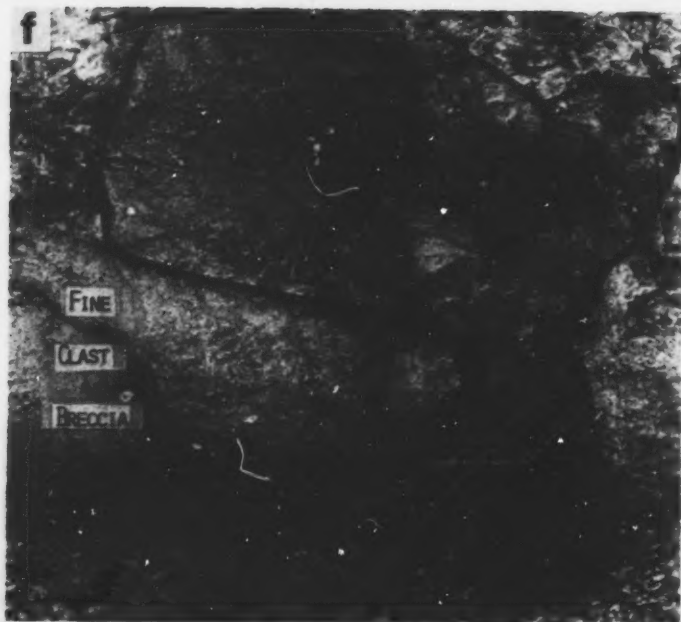


FIGURE 8f: View of basal 4 m of a 20-m-thick stratified breccia. Large breccia block (outlined) overlies a fine clast limestone breccia bed.

Upper Member, Unit "b"

Mechanism of Brecciation

Lithology

Unit "b" consists of interbedded limestone (in part breccia) and variously dolomitic limestone. It occurs only at Cape York in the interval 100 to 142 m above the base of the upper member. Unit "b" is similar in appearance to the lower member of the Cape Crauford Formation. Fossils in unit "b" at Cape York include brachiopods, echinoderm ossicles, stromatopoids, orthocones and corals. The vegetative parts of charophytes occur near the top of unit "b".

Basal contact

The basal contact of unit b is conformable at Cape York where it is marked by a 1.2-m-thick brecciated cherty limestone bed.

Distribution and Thickness

Unit "b" is absent because of erosion over most of the peninsula and was measured only in section H near Cape York (Fig. 2).

Age

A Silurian age was specified on the basis of conodonts processed for the present study (B. Chatterton, University of Alberta, pers. comm., 1985).

BRECCIA BODIES

The breccia bodies, which are most common in the Cape Crauford Formation, have various shapes and dimensions. Generally they can be grouped into discordant breccias that truncate bedding and stratified breccias that follow bedding.

Discordant Breccia Bodies

The vertical pipes of breccia include some that are up to 110 m high, 20 m wide at the base and 10 m wide at the top (Fig. 9a,9b). In the Port Bowen area, they are common and close enough together that they locally merge to form laterally continuous breccia bodies over 110 m wide and 80 m or more high. The pipes usually truncate the undisturbed strata abruptly, but in some cases strata bend sharply down and can be traced a short distance into the pipes. The pipes commonly extend down to a main stratified breccia, but isolated breccia pipes that do not bottom in stratified breccia also occur (Fig. 6). Some of the pipes and merged pipe bodies extend upward to the regional unconformity at the base of the upper member of the Cape Crauford Formation (Fig. 8e). None of the pipes cross the unconformity. The breccia clasts are usually angular to sub-angular and range up to 2 m in size. In some pipes there is an inverse grading of clast size.

Stratified Breccia

The individual stratified breccia units are laterally continuous for distances up to several kilometres, but are not traceable on a regional scale. Inverse size grading of clasts is common in the stratified breccias. A less common variant in the stratified breccias is a fine clast breccia a metre or more in thickness that is overlain by a coarse clast breccia (Fig. 8f).

The breccias of the Brodeur Group were interpreted, on the basis of reconnaissance work by Norris (1963) and Trettin (1965a, b, 1969), to have formed by the dissolution of evaporites. Evaporites are present in the Lower Silurian strata of north-eastern Somerset Island (about 70 km west of Brodeur Peninsula) where they are interpreted to have been deposited in a sabkha-type depositional environment (Reinson, 1978). In addition, a 9.1-m-thick bed of laterally extensive gypsum was found by Blackadar (1956) 97 m above the base of the formation near Baillarge Bay on the east coast of Admiralty Inlet (Fig. 1). In neither case, however, are the strata overlying the evaporites brecciated. It is common in many parts of the world to find evaporites in the subsurface represented by breccias at the surface. Unfortunately, Brodeur Peninsula is separated from the evaporites of Somerset Island and Borden Peninsula by Prince Regent and Admiralty inlets. Correlation of the Brodeur strata with the strata of Somerset Island is difficult because of rapid lateral facies changes. For these reasons hypotheses respecting the origin of the breccias is based on the field evidence gathered on Brodeur Peninsula itself.

Part of the problem lies in the fact that the process of evaporite-solution brecciation tends to destroy the very evidence needed for interpretation. Despite this, it should be possible, on the basis of the remaining undisturbed rock, to determine the original depositional environments and from this infer the presence of the evaporites. Such a scenario is not possible in the case of the Cape Crauford Formation on the Brodeur Peninsula because the available sedimentological evidence does not support the notion of a supratidal depositional setting akin to that in which the evaporites of the Cape Crauford Formation on Somerset Island developed. The Cape Crauford Formation of Brodeur Peninsula, especially in the lower part of the lower member, contains a diverse marine fauna (Fig. 4). Such a fauna, along with the lack of features indicative of a supratidal or even high intertidal depositional environment, is incompatible with the hypothesis that evaporites were once part of the sequence. If the argument that evaporites were not present is accepted, it follows that the brecciation cannot be attributed to evaporite-solution-collapse.

The available evidence suggests that the breccias of Brodeur Peninsula were formed by limestone dissolution. Between the lower and upper members of the Cape Crauford Formation of Brodeur Peninsula there is a regional unconformity that probably developed during a period of subaerial exposure. Significantly, many of the breccia bodies and collapse structures have their upper boundary coincident with the unconformity, suggesting that they were in some way related. Such a situation would be consistent with the development of a karst terrain and collapse of strata associated with the dissolution of limestone. This interpretation is supported by the presence of high-amplitude stylolites and breccias that occur preferentially in the non-dolomitic limestones. Further evidence for subaerial exposure includes the occurrence of polygonal desiccation cracks and a thick pisolite zone near the top of the lower member (Fig. 7). Also, if the sequence is considered as a whole, there is ample evidence for gradual shallowing to the top of the lower member of the Cape Crauford Formation.

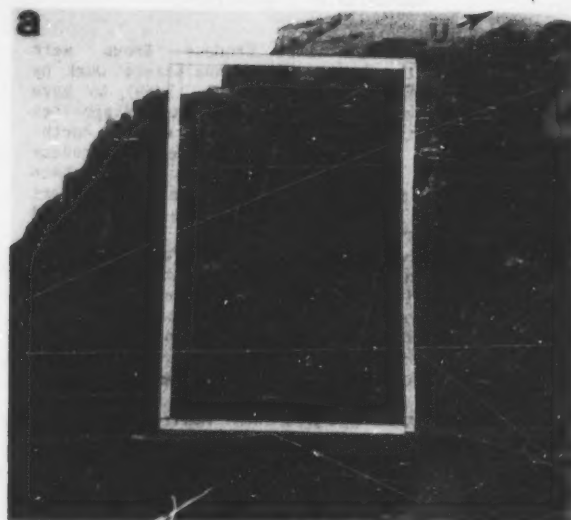


FIGURE 9a: Breccia pipes in the lower member of the Cape Crauford Formation just below the upper member (shown by the letter U), south shore of Port Bowen.

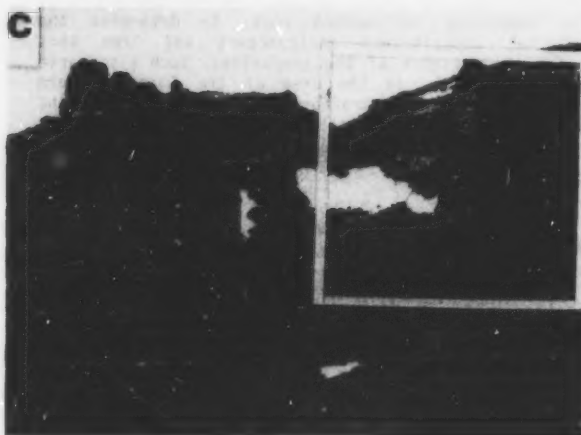


FIGURE 9c: View of breccia pipe 5 km southwest of Cape Crauford.



FIGURE 9b: Close-up view of two breccia pipes shown in the outlined area in Figure 8a.

While the bulk of the evidence points to the dissolution of limestone, there are still some unresolved problems. For example, some of the breccia clasts of Brodeur Peninsula display ductile deformation whereas most limestone karst terrains are developed on well-lithified carbonates. Resolution of these and other problems will require further research.

MINERAL POTENTIAL OF BRODEUR PENINSULA

A combination of factors makes the area extremely favourable from the standpoint of exploration for mineral deposits. Extensive near-surface Silurian

paleokarst breccia beneath an unconformity has been observed over an area of at least 5000 square kilometres. This area is situated between the Polaris and Nanisivik mines, which have combined published lead-zinc ore reserves of over 25 million tonnes (Gibbins, 1982). These reserves occur in solution caverns in carbonate rocks. Twelve of fourteen deposits that have been found in the Cornwallis Lead-Zinc District are in the Thumb Mountain Formation in brecciated dolostone (Gibbins, 1982), and according to a study of conodonts by Barnes (1973) this formation is only slightly older than the Baillarge. The mineralisation at the Truro Island and Polaris deposits is believed to have occurred during the Late Devonian (Gibbins, 1982) at a time when the Silurian



FIGURE 9d: Close-up view of breccia pipe outlined in Figure 9c.

paleokarst of Brodeur Peninsula was already developed.

In addition, over a dozen kimberlite pipes have been discovered on northern Somerset Island to the west of Brodeur Peninsula (Mitchell, 1976) and 35 to 40 have been found on Borden Peninsula to the east of the study area (D. Mason, Pan Canadian, oral comm., 1984). No kimberlite pipes have been found to date on Brodeur Peninsula perhaps because of the mantle of the cover sediments. The kimberlites on Somerset Island have a positive (500-2000 gamma) magnetic anomaly (Mitchell, 1976) so that pipes with similar anomalies should be readily identified by aeromagnetic surveys. The association of diamonds with some kimberlite pipes gives the area some potential in this regard.

CONCLUSIONS

This study permits the following conclusions:

1. The Brodeur Group is composed of the Baillarge and Cape Crauford formations, a succession of Ordovician and Silurian shelf carbonates.
2. The Baillarge Formation is informally divided into members A and B. Member A is about 140 m thick and consists of dolostone and interstratified limestone along with rare thin beds of shaly limestone and a thin laterally discontinuous solution breccia. Member B is 374 m thick and consists of fossiliferous, massive, resistant dolomitic limestone. A limestone solution breccia occurs 292 to 310 m above the base of the member.

3. The Cape Crauford Formation overlies the Baillarge Formation and is subdivided informally into lower and upper members.
4. The lower member is a shallowing-up sequence consisting of interstratified limestone, dolomitic limestone, calcareous dolostone and dolostone. Both stratified and pipe-solution breccias are present. The contact between the lower and upper members is an unconformity. Subaerial exposure is recorded in the rock by desiccation cracks, collapse structures and pisolites just below the unconformity.
5. The upper member is informally subdivided into units "a" and "b". Unit "a" is a distinctive orange-weathering massive dolostone that is the most mappable unit in the area. It attains a thickness of 110 m at Cape York. It is overlain by unit "b", a 42-m-thick sequence that consists of thin units of interstratified limestone (in part breccia) and variously dolomitic limestone.
6. Most of the breccias are believed to have formed by interstratal limestone solution during subaerial exposure between the time of deposition of the lower and upper members of the Cape Crauford Formation.
7. The lead-zinc mineral potential of the peninsula is believed to be heightened by the recognition of paleokarst.

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FIGURE 9e: Small collapse in the uppermost part of the lower member of the Cape Crauford Formation at section B, Jackson Inlet.

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DEVONIAN HARE INDIAN-RAMPARTS EVOLUTION,
MACKENZIE MOUNTAINS, N.W.T.:
BASIN-FILL AND PLATFORM-REEF DEVELOPMENT

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INTRODUCTION

The study area is in the Mackenzie Mountains approximately 110 km west of Norman Wells, N.W.T. (Fig. 1). Thirty-one stratigraphic sections through a Middle-Upper Devonian sequence consisting of the Hume, Hare Indian, Ramparts and Canol Formations have been examined in detail during three field seasons. A preliminary surface - subsurface study, jointly with Esso Resources Canada, was conducted at the conclusion of the 1984 field season (Muir et al., 1984). This combined study helped to define the geometric expression of facies and their mutual relationships in time and space.

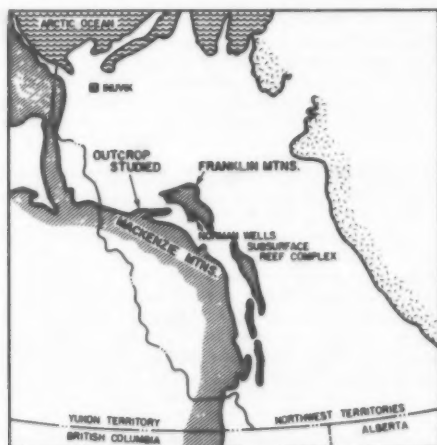


FIGURE 1: Location map of study area.

This paper summarizes our interpretation of Middle Devonian basin-fill and platform-reef development in the Mackenzie Mountains and Norman Wells areas.

STRATIGRAPHY

Relationships and relative ages of the lithostratigraphic units are shown in Fig. 2. The Hume and Ramparts Formations are predominantly limestone with minor shale interbeds. The Ramparts Formation is the surface equivalent of the Kee Scarp reservoir reef limestone at Norman Wells. The Hare Indian, Canol and Imperial Formations contain fine-grained terrigenous clastic lithofacies. Conformable relationships

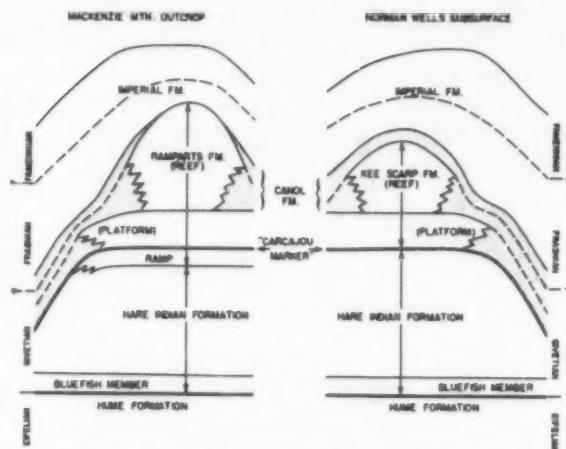


FIGURE 2: Stratigraphic terminology employed in this report.

purported here differ from other workers' interpretations. In particular, Warren and Stelck (1962), Bassett and Stout (1967), Gilbert (1973) and Braun (1966, 1977, and 1978) have interpreted a pre-Canol unconformity. However, intertonguing and onlapping relationships in both the surface and subsurface demonstrate that this is not the case in the Mackenzie Mountains-Norman Wells area (Figs. 3, 4).

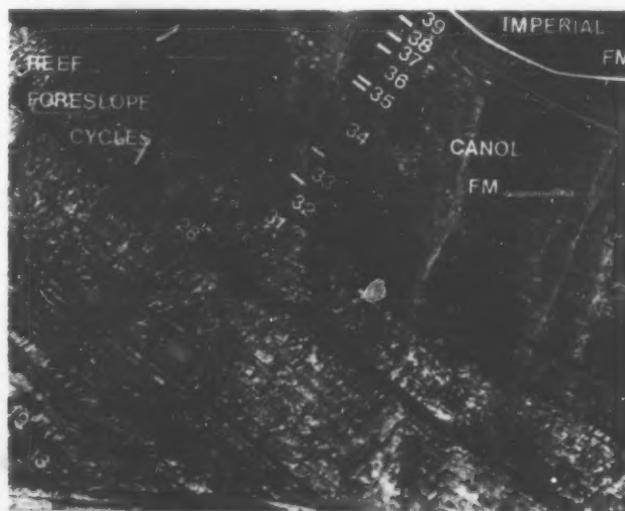


FIGURE 3: Reef foreslope and cycles at Powell Creek. Note interbedded black shales (units 30, 32).

Absence of faunal zones reported by earlier workers may simply reflect inadequate sampling through condensed sequences caused by very slow sedimentation and also possibly by infrequent submarine current erosion. Hence Canol shales in some areas rest on the basal black shales of the Bluefish Member of the Hare Indian Formation with no intervening Ramparts Formation or upper Hare Indian Formation. This is attributed to depositional thinning down a clinoform surface into a deep-water basin.

BASIN-FILL AND PLATFORM-REEF EVOLUTION

The Hare Indian and Ramparts sequences have a remarkably similar style of evolution in both the Mackenzie Mountains and the Norman Wells subsurface. Examination of these deposits shows that they consist of two major or first-order shallowing-upward cycles of sedimentation:

- 1) Lower first-order cycle - includes the entire Hare Indian Formation and the lower "ramp" unit of the Ramparts Formation. Maximum thickness in the study area is 200m.
- 2) Upper first-order cycle - includes the remaining portion of the Ramparts Formation, the Canol shales and the Imperial shales and sandstones. The top of this shallowing-upward sequence was not ascertained.

The use of these and lower-order cycles permits the definition and delineation of true time stratigraphic units. Well-developed second-order shallowing-upward cycles (avg. 20-25m thick) are present in the Mackenzie Mountains and Norman Wells Ramparts reef complexes. The cycles can be correlated from one complex to the other and thus provide a more detailed

stratigraphic framework for the Ramparts Formation. This cyclicity likely resulted from relative changes in sea level involving pronounced eustatic sea-level rises. Many cycle tops have been correlated from the Mackenzie Mountains buildup to the Norman Wells buildup on the basis of their position relative to three conspicuous markers (Fig. 4):

- 1) top of the Hume platform.
- 2) base of a regionally extensive shale unit known as the "Carcajou Marker".
- 3) base of reef cycle 4 which represents a significant backstep of the reef margin in both complexes.

Pronounced sea-level rises caused abrupt facies shifts and formation of these markers. Cross-section comparisons through both complexes show that only the subsurface cycles with the greatest lateral facies shift can be recognized in the Mackenzie Mountains complex. The change from one second-order cycle to a succeeding cycle is marked by an abrupt change to deeper-water facies.

The Mackenzie Mountains buildup culminates with two cycles of reef growth (cycles 5 and 6 in Fig. 4) that are not present in the subsurface (Muir et al., 1984). These two additional cycles account for the appreciably thicker reef section in the Mackenzie Mountains.

Detailed sedimentology of the Hare Indian and Ramparts Formations is discussed in Dixon (1984), Muir and Dixon (1984), and Muir et al. (1984). The remainder of this paper discusses the evolution of the Hare Indian-Ramparts sequence in terms of growth stages related to episodic rises in sea level (Fig. 5). For ease of cross-reference, the Norman Wells reservoir zonation scheme (K1A to K4) employed by Esso Resources Canada is applied to the Mackenzie Mountains buildup. Muir et al. (1984) provide a more complete discussion of this zonation format.

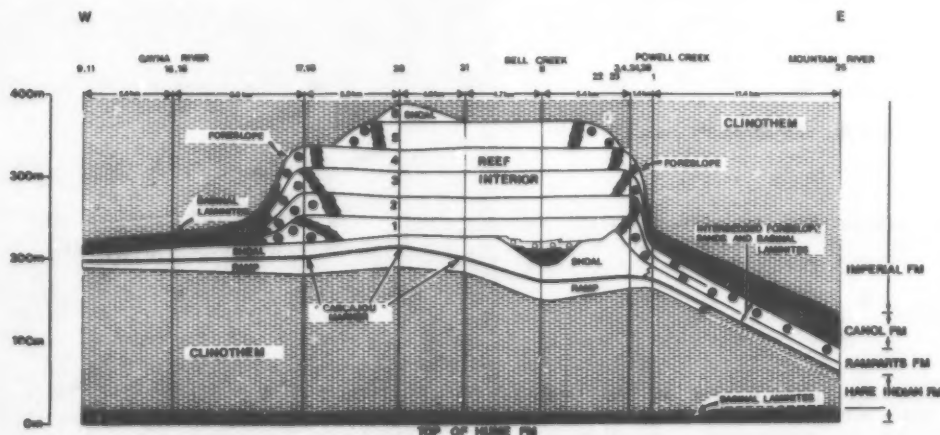


FIGURE 4: Schematic drawing of Mackenzie Mountains reef complex. Six second-order reef cycles are present.

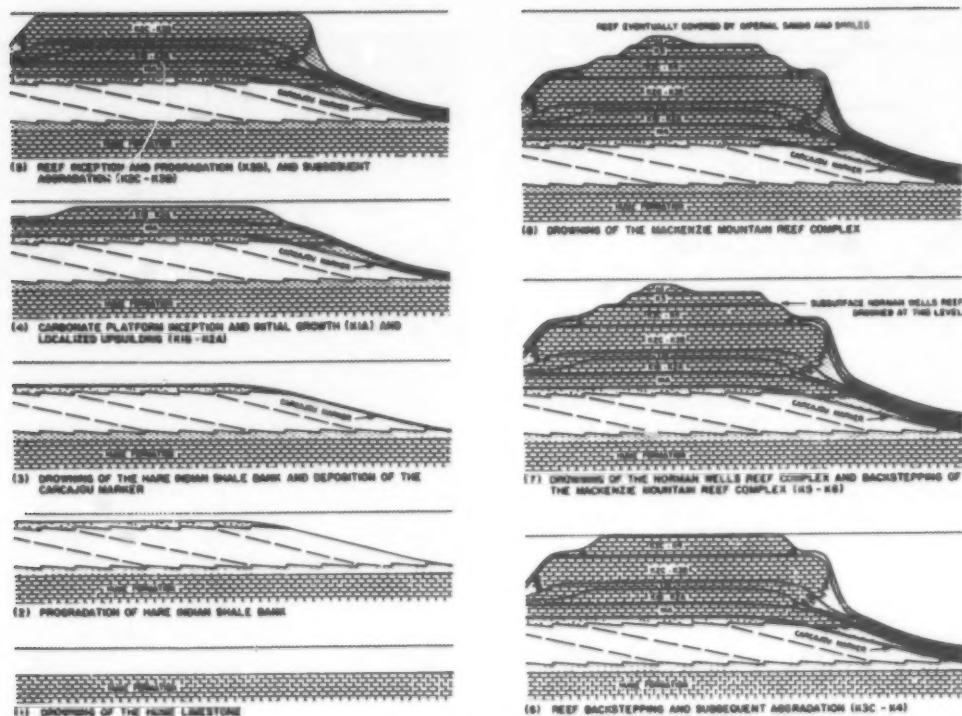


FIGURE 5: Evolution of the Hare Indian clastic wedge and the Ramparts reef complex. Brick pattern represents shallow water carbonates. Stippled pattern indicates shallow water ramp facies. Deep water shales and limestones shown by broken line pattern.

Stage 1 A pronounced and rapid rise in sea level occurred near the beginning of Givetian time. The Hume platform was drowned and sea level continued to rise until water depths in excess of 150m were achieved in the Mackenzie Mountains-Norman Wells area.

Stage 2 Progradation of the Hare Indian clastic wedge into this deep-water basin accompanied a slowing of sea-level rise. Sedimentation occurred on a clinoform surface both from hemipelagic suspension and low density turbidity currents. Slow suspension deposition on the basin floor resulted in widespread thin accumulation of black, organic-rich shales of the Bluefish Member. As progradation continued, the areas of the present day Mackenzie Mountains and the Norman Wells field built closer to wave base and became sites of *in situ* sedimentation on a relatively shallow-water ramp. Mixed carbonate-siliciclastic deposition tended to be restricted to Hare Indian paleotopographic highs, while less diluted carbonate deposition occurred off-lobe and in paleotopographic lows (Fig. 6).

The entire Hare Indian Formation and ramp portion of the Ramparts Formation represents a coarsening-upward sequence in terms of both bed thickness and grain size. Cyclic occurrence of shales and limestones is best observed slightly off-lobe in deeper waters where lithological contrasts are more evident (i.e. zone of mud accumulation). These second-order cycles are of the same magnitude as those observed in the Ramparts reef complex (average thickness 15-

25m). The cycles are characterized by a shaly lower portion and a limestone-dominated upper portion (Fig. 7).

The ramp succession varies in thickness from less than 10m in off-reef sections to 26m under the reef. The lower portion of the ramp consists of bioturbated shales, nodular lime mudstones and wackestones. The upper ramp succession is a heterogeneous mixture of facies (including carbonate bindstones, rudstones, floatstones and packstones) with a significant siliciclastic component expressed as storm deposits particularly towards the top of the succession.

Stage 3 This stage began with drowning of the ramp succession of the Ramparts Formation and deposition of the Carcajou Marker shale lithofacies. Rapid sea-level rise was accompanied by decreased sediment supply and arrested clastic wedge progradation in both the Mackenzie Mountains and Norman Wells areas. This marked the termination of the first basin-fill cycle or lower first-order cycle.

Stage 4 Following the abrupt drowning of the ramps, a branching coral fauna became established on topographic highs. Corals built toward sea level and in some locations gradually became replaced by cylindrical stromatoporoids in shallower waters. The resulting carbonate platform lithofacies built basinward, filling in lows and creating broad but still submerged platforms by the end of K1A time. The development of the carbonate platform was cyclic. A



FIGURE 6: Distribution of siliciclastic-dominated ramp facies (stippled pattern) and carbonate-dominated ramp facies (brick pattern). Underlying Hare Indian isopach data show that carbonate ramp facies developed in paleotopographic lows and away from major terrigenous clastic input sites. Type locality for the Ramparts Formation is shown. Isopach data (m) taken from Williams (1980, written communication); Muir and Dizon (1984); Pugh (1983).

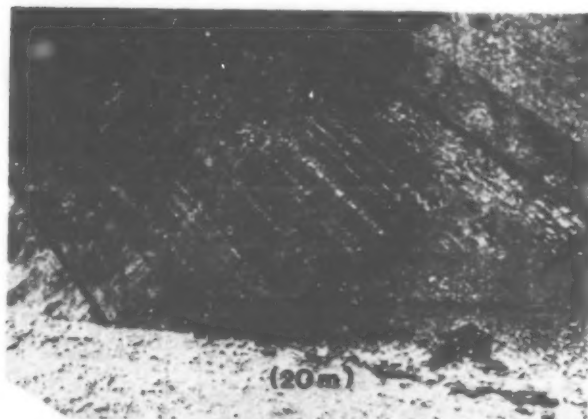


FIGURE 7: Coarsening- and thickening-upward sequence in oolite facies, Hare Indian Formation. Alternation of dark green-grey calcareous shales and argillaceous calcisiltites. Section 7, units 004 to 008 (arrows) approximately 20m thick.

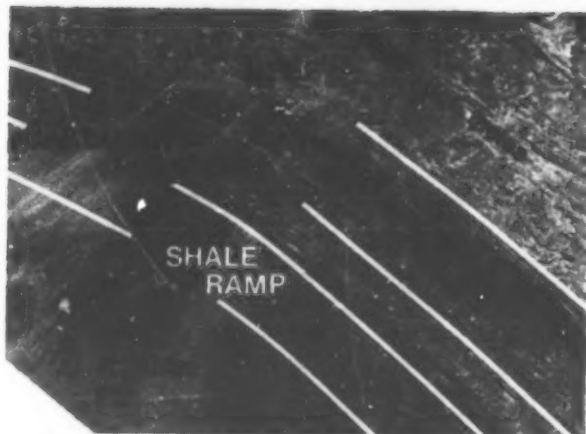


FIGURE 8: Carbonate platform at Bell Creek (section 8) composed mainly of allochthonous foreslope facies. Overlain by embayment facies of first reef cycle (approximately 31m thick).

subsequent acceleration of sea-level rise led to relatively deep water conditions and deposition of dark bituminous shales and foreslope limestones on slightly lower areas of the carbonate platform (Fig. 8). However, on more positive areas of the platform a stromatoporoid-coral fauna continued to grow and keep pace with the rise of sea level. These conditions led to the creation of more areally restricted shoals on the carbonate platform with surrounding deeper water areas during K1B-K2A time (Fig. 4).

Stage 5 Reef cycle 1 inception and progradation (K2B) and subsequent aggradation of reef cycles 2 and 3 (K2C through K3B) characterize stage 5.

A sea-level rise terminated carbonate platform growth and reef-forming benthos colonized the margins of previous platform highs. During K2B time, the resulting reefs built very slowly upwards but prograded laterally and filled adjacent lows. The reef margins sheltered an interior lagoon where *Amphipora* wackestones and tidal-flat facies were deposited.

Successive sea-level rises between K2B and K3B time produced only minor backstepping of the reef margins and the reefs grew essentially upwards in the same general positions. Shallow water lagoonal and tidal-flat deposits continued to form in the sheltered interiors of both the Norman Wells and Mackenzie Mountains reefs.

Fig. 3 displays the lowermost three reef foreslope sand cycles (K2B, K2C-D, K3A-B) exposed at Powell Creek ("Allochthonous Beds" of MacKenzie, 1970).

Stage 6 An acceleration in sea-level rise following stage 5 led to substantial backstepping of the reef margin and onlapping of Canol basinal shales during stage 6, as illustrated in Fig. 5. Following sea-level rise, the reef margins again accreted upwards (K3C-K4) with only small increments of backstepping in the Norman Wells reef complex. This could

not be resolved at outcrop scale in the Mackenzie Mountains buildup. In any case, the rapid sea-level rise initiating stage 6 resulted in a change from restricted lagoonal to more open marine sedimentation in both the Norman Wells and Mackenzie Mountains buildups.

Stage 7 Another pulse of accelerated sea-level rise drowned the Norman Wells buildup and resulted in backstepping of the Mackenzie Mountains reef margin (cycles 5 and 6, i.e. K5, K6). During K5 time in the Mackenzie Mountains area, open-marine, slightly deeper water lagoonal sedimentation continued in the reef interior (Fig. 9). A deepening at the beginning of K6 time led to the formation of an areally restricted cylindrical stromatoporoid-coral shoal lacking reef margins. Open marine conditions prevailed.

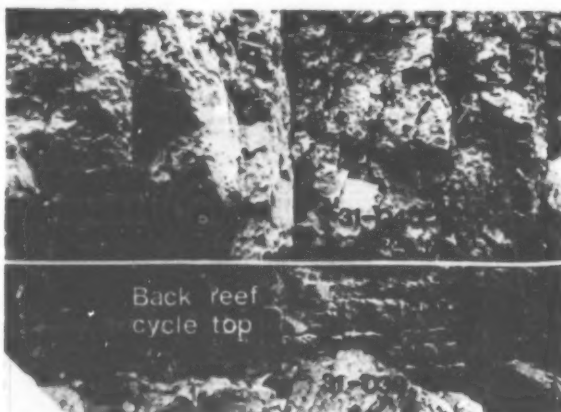


FIGURE 9: Dark, thin-bedded *Amphipora* wackestones abruptly overlain by back reef sediments (31-040, base of K6 cycle) showing a more open marine aspect (atrypid brachipods, robust cylindrical stromatoporoids and albus stromatoporoids). Scale is 3m long. Section 31, units 39, 40.

Stage 8 A rapid rise in sea level ultimately caused the drowning of the Mackenzie Mountains reef complex. Following cessation of reef growth, both complexes remained uncovered by sediments for a considerable period of time. Eventually prograding Imperial clinothem and Canol fondothem sediments downlapped and overlapped the reef complexes, finally covering them. There is no evidence for subaerial exposure.

DISCUSSION

Study of the Hare Indian and Ramparts formations reveals that distinct cycles of sedimentation can be correlated between the Mackenzie Mountains and Norman Wells buildups. Successful correlation of cycles between two widely separated complexes suggests that similar procedures applied elsewhere may aid in understanding the evolution of other carbonate platforms. Furthermore, surface studies can be a valuable complement to subsurface studies, in the following ways:

1) Proper recognition of facies and interpretation of their manner of deposition are sometimes difficult in subsurface studies. However, in surface studies careful mapping and walking out of beds permit the establishment of true facies relationships. This is particularly useful at buildup margins. More efficient field development may be undertaken through determination of reservoir continuity and correlation of dense permeability barriers. This added knowledge for field development may result in field extensions and well deepenings.

2) Field studies also provide a much improved means of studying non-reservoir units such as the Hare Indian Formation. The Hare Indian clastic wedges functioned as topographic highs upon which the Ramparts carbonate lithofacies were initiated.

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PRELIMINARY REPORT ON THE GEOLOGY
OF THE WILSON ISLAND GROUP
GREAT SLAVE LAKE, N.W.T.

by

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INTRODUCTION

The early Proterozoic Wilson Island Group is a succession of metasedimentary and metavolcanic rocks exposed in a northeast-trending belt in the East Arm of Great Slave Lake, Northwest Territories. Major exposures occur on the Outpost Islands, on Wilson Island, and in the Basile Bay area. The latter two areas lie on opposite sides of the McDonald-Wilson fault. The Wilson Island Group has been mapped and studied stratigraphically by Stockwell (1936), Reinhardt (1969), and Hoffman *et al.* (1977). Paleocurrent and geochemical analyses have been reported by Yeo (1976).

Although the Wilson Island Group is completely fault bounded (Bowring *et al.*, 1984), it is inferred to rest unconformably on Archean crystalline basement; clasts of Wilson Island Group metasediments contained in conglomerates near the base of the less-metamorphosed Sosan Group indicate that another unconformity separates the Wilson Island Group from the younger Sosan Group (Hoffman *et al.*, 1977).

This report summarizes results of preliminary mapping completed during the 1984 field season.

STRATIGRAPHY AND SEDIMENTATION: WILSON ISLAND AREA

The most detailed mapping was carried out across the southwestern part of Wilson Island and on adjacent islands, where the oldest rocks of the section are exposed (Fig. 1). Massive amygdaloidal basalt flows predominate in the lower 200 to 300 metres of exposed section (Fig. 2). Flow breccias, intercalated mudstone and mafic ashstone form about 20 percent of this section. The basalts are overlain by about 600 metres of interstratified immature arkose, volcaniclastic rocks, and several amygdaloidal basalt flows. Most of the flows are laterally continuous, providing excellent markers for mapping. Volcanic breccias and tuffs contain blocks and bombs up to 20 cm across, and are possibly of intermediate composition.

The arkoses are moderately to poorly sorted, and chiefly medium to fine grained. They typically form beds less than 50 cm thick, with sharp, scoured bases, and are interbedded with thinner layers of very fine-grained sandstone and siltstone. The arenites are extensively trough cross bedded, whereas the siltier layers display parallel and ripple laminations. These features collectively are typical

SOSAN GROUP

Sh HORNBY CHANNEL Fk: Sandstone

Inferred unconformity

WILSON ISLAND GROUP

Dacite

Conglomerate

Sandstone

Basalt

Conglomerate, sandstone

Basalt

Basalt

Sandstone, siltstone, volcaniclastics, minor basalt

Basalt, minor mudstone

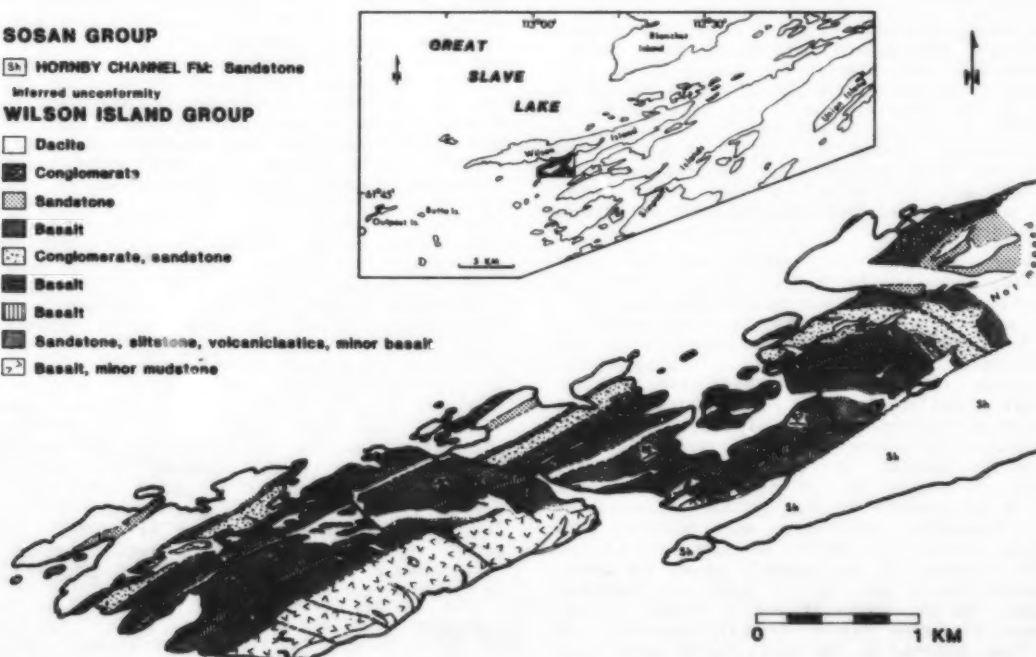


FIGURE 1: Geologic map of southwest end of Wilson Island.

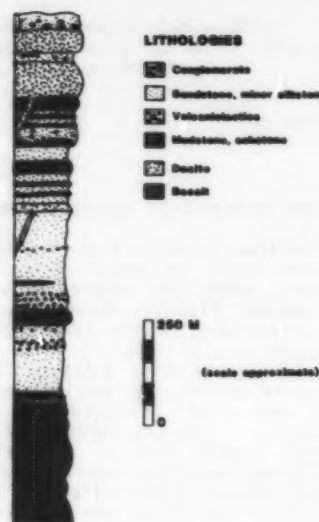


FIGURE 2: Schematic stratigraphic section, south-western Wilson Island.

of distal braided river and alluvial plain deposits (Rust, 1978). Ball-and-pillow structures and other soft-sediment deformation structures are abundant in the arkoses, indicating rapid sedimentation in an unstable environment. Contemporaneous volcanism provides additional evidence for syndepositional tectonic activity.

The arkoses are sharply overlain by conglomerates that contain pebbles and cobbles of vein quartz, metasediments, and granitoid rocks. These conglomerates grade upward into pebbly arkose and arkose similar to the underlying unit. If, as suggested by Yeo (1976), the Wilson Island Group was deposited in a fault-bounded trough or rift, syndepositional faulting may have periodically modified both the paleoslope and the relief of the source area, causing sudden changes from sand-dominated to gravel-dominated deposits.

Another argument in support of the rift hypothesis, suggested by Hoffman *et al.* (1977), is the coexistence of mafic and felsic components of a bimodal suite of volcanic rocks. The conglomerate is capped by a basalt flow, in turn overlain by two semiconcordant sheet-like bodies of plagioclase-phyric dacite (the name dacite is here based upon the phenocryst assemblage in these rocks, which in almost all instances consists entirely of plagioclase) that are separated by arkose and minor volcaniclastics. The dacites consist of plagioclase phenocrysts (commonly in excess of 20 percent of the rock volume) in an aphanitic to fine-grained groundmass of quartz, muscovite, chlorite, and biotite. The plagioclase phenocrysts exhibit Carlsbad, albite, and pericline twinning and are commonly zoned. The sheet-like dacite layers are fed by dykes of identical mineralogy. On the southwest end of Wilson Island, the lower of these layers directly overlies the basalt flow, whereas to the east, in the axial zone of a major anticlinorium (near the eastern edge of Fig. 1), at least 15 to 20 m of arkose separates a similar felsic body from the basalt. Although the contact of this body with arkoses and conglomerates to the north was not observed, it is interpreted as a

discordant intrusion. Until further mapping has been completed, it remains unclear whether any of the felsic rocks of the Wilson Island Group are truly extrusive. Thus, a U-Pb zircon age of 1928 ± 11 Ma determined from these rocks (Bowring *et al.*, 1984) must be considered a minimum age of the Wilson Island Group.

Further up section to the north and east of the area shown in Fig. 1 is more arkose, basalt, and a thick, generally structureless paraconglomerate containing abundant pink felsite clasts. The clasts may have been derived from the underlying dacites. These massive, matrix-supported conglomerates are tentatively interpreted as debris-flow deposits.

On northern Wilson Island the conglomerates give way to coarse arkosic arenite, which is overlain by at least 1200 m of intercalated fine- to medium-grained arkose, subarkose, dolomitic subarkose, and sandy dolostone. A few metres of rhythmically interlayered sandstone and mudstone generally characterize the transition from sandstones to dolostones. The sandstones on northern Wilson Island have been interpreted by Yeo (1976) to be braided stream deposits. The intercalated dolostones could represent playa lake deposits, but more likely mark periods of marine transgression. This is supported by close association of the dolostones with rhythmically interbedded sandstone and mudstone, a common facies in tidal deposits (Reineck and Singh, 1973).

STRATIGRAPHY AND SEDIMENTATION: BASILE BAY AREA

On Basile Peninsula (80 km northeast of the Wilson Island area mapped in detail, Fig. 1), arkoses and subarkoses similar to those on northern Wilson Island are overlain by massive to laminated dolostone. Sharply overlying the dolostone is about 500 m of pink, fine-grained arkosic to subarkosic arenite displaying trough and planar cross bedding and heavy-mineral laminations. These may be of fluvial origin as suggested by Yeo (1976), but in part could represent shallow marine deposits. The arenites grade upward into dark gray, parallel-laminated to ripple-laminated siltstone and mudstone through a transitional facies of intercalated arenite, siltstone, and mudstone. Arenites in the transitional facies exhibit low-angle cross bedding and minor herringbone cross bedding; wavy to lenticular-bedded mudstone also is abundant. The transitional facies was probably deposited in a tidal environment, whereas the overlying siltstones and mudstones are probably deeper-water deposits. The siltstones and mudstones are overlain by white quartz-arenite granulestone, feldspathic sandstone and siltstone with extensive dark gray iron-oxide-rich layering, and very fine-grained arkosic arenite/mudstone cycles. The cycles consist of sharp-based sandstones with graded bedding and rare cross bedding, parallel-laminated muddy siltstone, and mudstone. Individual fining-upward cycles 5 to 100 cm thick form a succession which is several hundred metres thick, overlain by dark green mudstones. The fining-upward cycles are tentatively interpreted to be turbidity current deposits.

STRUCTURE

In the Wilson Island area, the major structure is a northeast-trending anticlinorium (Fig. 1). The anticlinorium has been breached by a fault which juxtaposes clastic rocks of the younger Sosan Group against the core zone and northwest-facing limb of

the fold. Open, inclined/overtaken folds with southeast-dipping axial surfaces plunge gently to steeply northeastward. Southeast-dipping foliation and south-southwest-plunging stretching lineations are extensively developed in the volcanic rocks. Northwest-trending faults in the lower part of the section generally show right-lateral separation (up to 100 m). Some of these faults appear to have been active before or during emplacement of the felsic intrusions.

On Basile Peninsula, most of the rocks dip steeply to the southeast. Tight to isoclinal, overturned folds plunge moderately eastward. Axial-plane cleavage is subparallel to bedding on the fold limbs. In the Basile Bay area, the Wilson Island Group is in fault contact with mudstones and volcanic rocks of the Kahochella Group to the northwest, and is bounded on the southeast by the McDonald-Wilson fault. The McDonald-Wilson fault and related northwest-trending dextral transcurrent faults are the youngest structures. North-northeast-trending dextral faults in the Basile Bay area and the northwest-trending faults on Wilson Island may be related to the McDonald fault system, or they may be older.

OUTPOST ISLANDS: PETROLOGY AND ECONOMIC GEOLOGY

Metamorphism of the Wilson Island Group attains its highest grade at the Outpost Islands. A sample of quartzite collected from the Outpost Islands contains staurolite and garnet porphyroblasts, and biotite which has in part been retrogressively metamorphosed to chlorite.

Shear zones in clastic rocks on the Outpost Islands have been mined for gold, tungsten, copper, silver, and tin. The possibility that the Wilson Island Group was deposited in a rift makes it a prime target for various precious and base metal deposits. Massive sulfides may be associated with the mafic lavas, and the carbonates may host Mississippi Valley-type lead and zinc sulfides. Small gossans are common in the basalts. Both the basalts and the dolostones exhibit copper staining in several localities.

CONCLUSIONS

The oldest exposed rocks of the Wilson Island Group are intercalated arkose, basalt flows, felsic intrusions and possibly felsic flows. The depositional basin was probably a fault-bounded trough or rift. Tectonic activity during the early stages of sedimentation is indicated by paraconglomerates, soft-sediment deformation structures, and volcanics. The arkoses were probably deposited by braided streams. Higher in the section, arkoses are intercalated with dolostones, which may represent periodic marine transgression. A succession of fine-grained arkoses, siltstones, and mudstones overlies the dolostones. These were probably deposited in various fluvial, marginal marine, and deep water settings. Rocks of the Wilson Island Group have been metamorphosed (greenschist to amphibolite facies) and compressed into eastward- to northeastward-plunging, open to isoclinal folds, and subsequently dissected by dextral transcurrent faults of the McDonald fault system.

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PROGRESS REPORT ON MAPPING
GREAT BEAR MAGMATIC ZONE, WOPWAY OROGEN

by

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The 1984 field season was spent mapping within the north-central portion of the Great Bear Magmatic Zone, Wopway Orogen. Earlier (1:125,000) mapping of the Great Bear Zone indicated three major subdivisions of supracrustal rocks: the LaBine, Sloan and Dumas Groups (Hoffman, 1978). The total exposed thicknesses of these units exceeds 35 km, clearly not a true stratigraphic thickness. Thus, a major effort was initiated in the summer of 1984 to map a transect across the Great Bear Magmatic Zone in order to better understand stratigraphic and structural relationships. This work, in conjunction with earlier work by Hildebrand (1983) and Bowring (1981) to the west and east, respectively, will allow construction of geologic cross sections and a better understanding of the magmatic and structural history of the zone. This report represents a progress report and many of the conclusions are preliminary; at least one additional field season will be required to complete the transect. Geologic maps (1:50,000) are in preparation as are chemical and petrographic studies.

The first two thirds of the summer was spent mapping the area around McClaren and Junius Lakes (86K/10, 86K/7). The Sloan Group is well exposed in this area and consists of a thick sequence of east-dipping dacitic and rhyolitic ash-flow tuffs intruded by porphyries of similar composition. A distinctive, thick, ash-flow tuff unit outcrops in this area and is characterized by abundant flattened clots (up to 10 cm long) interpreted as unvesiculated magma incorporated during ash-flow tuff eruption. This ash flow contains very little pumice, but when observed it defines a foliation parallel to that of the flattened magma clots. The ash flow is overlain by a distinctive interval of welded ash-flow tuffs and interbedded volcanoclastic sedimentary rocks. The dacitic ash flows are underlain by a lithic-rich rhyolitic ash flow that is well exposed at the west end of Junius Lake. Combined stratigraphic thickness for this interval is approximately 6 km. The thick dacite tuff is intruded by large sills of plagioclase-hornblende porphyry (up to 0.5 km) that are difficult to distinguish from the ash flows except that they lack lithic fragments, magma clots, and in general the phenocrysts are more euhedral than in the ash flows. The flattened clots in the ash-flow tuffs are identical to the porphyry material and porphyries are interpreted as being comagmatic with the ash flows.

A distinctive, brick-red, densely-welded, and lineated rhyolitic ash-flow tuff overlies the sequence of bedded tuffs and volcanoclastic sedimentary rocks. This is in turn overlain by a sequence of pumice-poor dacitic ash-flow tuff intruded by dikes and blob-like bodies of crystal-poor rhyolite porphyry. At the northeast end of Junius Lake, the sequence described above is juxtaposed against a thick west-dipping sequence of dacitic to andesitic lavas and tuffs along a peculiar east-west trending fault zone. The fault zone is characterized by a well-developed protomylonitic fabric. Along the fault zone, compaction foliation in an ash-flow tuff has been rotated into parallelism with the fault zone with a stretching lineation

parallel to the trend of the fault. The deformation clearly predates a large quartz monzonite to monzonite pluton that intrudes the volcanic rocks and the fault zone. The sequence of lavas and tuffs south of the fault zone has not been recognized to the north or south, and until further mapping is completed, we won't know where it fits into the regional stratigraphy. Similar fault zones have been recognized during reconnaissance to the southwest, although further work is needed to fully understand the kinematic significance of this style of faulting.

The last part of the summer was spent west of Junius Lake in the vicinity of "915" Lake where preliminary evidence suggests that the area is within an ash-flow tuff cauldron. A large, flow-banded, rhyolite dome previously described by Hoffman & McGlynn (1977) intrudes siltstone and mudstone and has an intense alteration halo associated with it. The alteration is dominated by intense silicification and bleaching of surrounding ash-flow tuffs. A distinctive, densely welded, flow-banded ash-flow tuff crops out west of "915" Lake and has large (up to 100 meter) blocks of siltstone "floating" within it. Further work is needed here, but the association of a thick dense ash-flow tuff containing large megabreccia blocks that must have fallen into the tuff as it was accumulating are suggestive of a cauldron environment.

The volcanic rocks of the Sloan Group have been intruded by two types of granitic pluton. The older plutons are quartz monzonite to monzonite in composition and have been folded along with the volcanics about NW-trending fold axes. Following folding, coarse-grained syenogranite plutons were intruded as sheet-like bodies. Preceding but probably associated with the syenogranite plutons are a swarm of granitic dikes that are well exposed in the area between McClaren and Junius Lakes. They range in composition from quartz monzonite to syenogranite and are in general subparallel to the NE trend of younger transcurrent faults. This suggests that the area was undergoing NW-SE extension following folding of the Great Bear Magmatic Zone.

The NE-trending right-lateral transcurrent faults are the youngest tectonic event recognized in the area and generally have less than 5 km of separation along them. Significant mineralization is associated with one of the faults at the NE end of McClaren Lake and it has been prospected extensively. Large quartz veins and stockworks intrude both volcanic and granitic rocks and mineralization is characterized by massive pods of chalcopyrite and bornite. In addition, significant molybdenite was noted, although this has not been described in the assessment files.

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